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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CLAY MINERALOGY OF MISSISSIPPI RIVER DELTAIC SEDIMENTS

BY M. S. TAGGART, JR., AND A. D. KAISER, JR.

ABSTRACT

Late Quaternary Mississippi River deltaic sediments off the coast of Louisiana were studied to obtain information on possible clay-mineral alteration in a marine environment. The samples utilized were cores from four foundation test borings and included clays which had been deposited in several environments. For comparison, studies were also made of the clay minerals transported by the Mississippi and Red rivers.

The results support the conclusion that clay minerals do not alter appreciably as a result of diagenesis in a marine environment and that the clay-mineral composition of source sediments determines the clay-mineral content of marine sedimentary rocks. All the deltaic samples contained montomorillonite, illite, kaolinite, and chlorite. The most noticeable variation among the samples was in montmorillonite content. Montmorillonite was the most abundant clay mineral in most of the deltaic samples but was much less concentrated in some of the remaining samples. Variations in montmorillonite content did not correspond with differences in depositional environment. Clays with high montmorillonite content occurred in both essentially marine and fresh-water facies.

Analyses of samples of Mississippi River and Red River sediments indicated that these rivers carry the same suite of clay minerals as is present in deltaic sediments. Furthermore, samples of Red River sediment, characteristically reddish-brown, were predominantly low in montmorillonite content, whereas samples of the gray Mississippi River sediment were high in montmorillonite content. These differences in clay-mineral composition and color correlate substantially with corresponding variations in samples of deltaic sediment.

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INTRODUCTION

Clay minerals make up a large fraction of sedimentary rocks. Although the origin of clay minerals is still largely unknown, progress in understanding the problem has been facilitated in recent years by the adaptation of X-ray diffraction techniques to the examination of clays. These techniques permit identification and classification of clay minerals by means of their crystal structures.

That clay minerals form during weathering of igneous rocks is generally recognized. It might be assumed that the problem of the origin of clay minerals is essentially one of how these minerals form during weathering, were it not for the possibility of later diagenesis. It has been suggested that clay minerals are sensitive to environmental changes and may be unstable in certain environments. In a particular environment, unstable clay minerals may alter to clay minerals stable in that environment. Grim (1958) recently marshaled the evidence in favor of this "diagenesis concept." Environmental influences, according to the concept, determine to a large extent the distribution of clay minerals in sedimentary beds.

The possibility of clay-mineral alteration in marine waters has been widely entertained because of the importance of the marine environment in the depositional history of sedimentary beds. Grim and Johns (1954), from work with sediments in San Antonio Bay, off the Texas coast, concluded that montmorillonite alters to chlorite and illite in saline waters. Johns and Grim (1958), studying sediments of the Mississippi River delta, concluded that chlorite and illite are formed from montmorillonite in that area. Powers (1957) has suggested that illite alters to chlorite in the saline waters of the Chesapeake Bay area. These researches suggest that illite is an intermediate product and chlorite is the stable end-product of marine diagenesis.

The subject of marine diagenesis of clays, however, has become quite controversial. Accumulating evidence indicates that clays may not respond significantly to the normal marine environment. Abundant montmorillonite has been found in deep-water sediments of some oceanic basins (Murray and Harrison, 1956) and in ancient marine beds (Simons and Taggart, 1954; Weaver, 1958). In addition, Milne and Earley (1958) failed to find evidence for the genesis of chlorite in clays brought to the northeast Gulf of Mexico by the Mississippi River and by other rivers, although they found evidence of the formation of illite where sedimentation was very slow.

Further progress in understanding the fate of clay minerals in a marine environment requires (1) greater attention to the geology of the areas studied and (2) resolution of uncertainties in the interpretation of X-ray diffraction data. The principal evidence offered in support of marine diagenesis is the observation of changes in clay-mineral content from fresh- to salty-water sediments in bays and estuaries. These observations have little significance unless accompanied by strong evidence that other factors, such as varying source or fractionation by sedimentation, have not contributed appreciably to the observed changes. Determination of the relative importance of these factors in a particular area requires information on all the geologic processes which may have influenced clay-mineral distribution in that area.

Much information is available on the geological history of the Mississippi River and its deltaic deposits. Fisk and McFarlan (1955) have made extensive studies of the region; their results are based on examination of sediments from hundreds of boreholes and on a chronological framework established by radiocarbon dating (Brannon *et al.*, 1957). The results of their work make possible the comparison of clay-mineral assemblages revealed by X-ray diffraction techniques with environments of deposition and with varying source of sediment. This geological information also permits study of the possible effects of fractionation during transport away from the river mouth and of possible changes with time after burial.

In regard to uncertainties in the interpretation of X-ray diffraction data, it should be noted that conflicting analyses of Mississippi River sediments have been reported. Johns and Grim (1958) concluded that these sediments contained montmorillonite, illite, and chlorite, but that kaolinite, if present, was very minor. On the other hand, Milne and Earley (1958) reported montmorillonite, kaolinite, and illite, but no chlorite. It seems likely that this discrepancy arose from differences in the interpretation of X-ray patterns rather than from differences in the patterns themselves. Unfortunately, the original patterns were not published. An important part of the present investigation, therefore, was examination of X-ray data for evidence of different clay minerals. The report includes interpretation of these data and also illustrations of the X-ray diffraction patterns.

Geological History of the Sediments

The Late Quaternary deltaic sediments of the Mississippi River were deposited following the peak of the last glaciation. At the time of the last glacial maximum, approximately 40,000 years ago, sea level was about 450 feet lower than at present, and the Mississippi

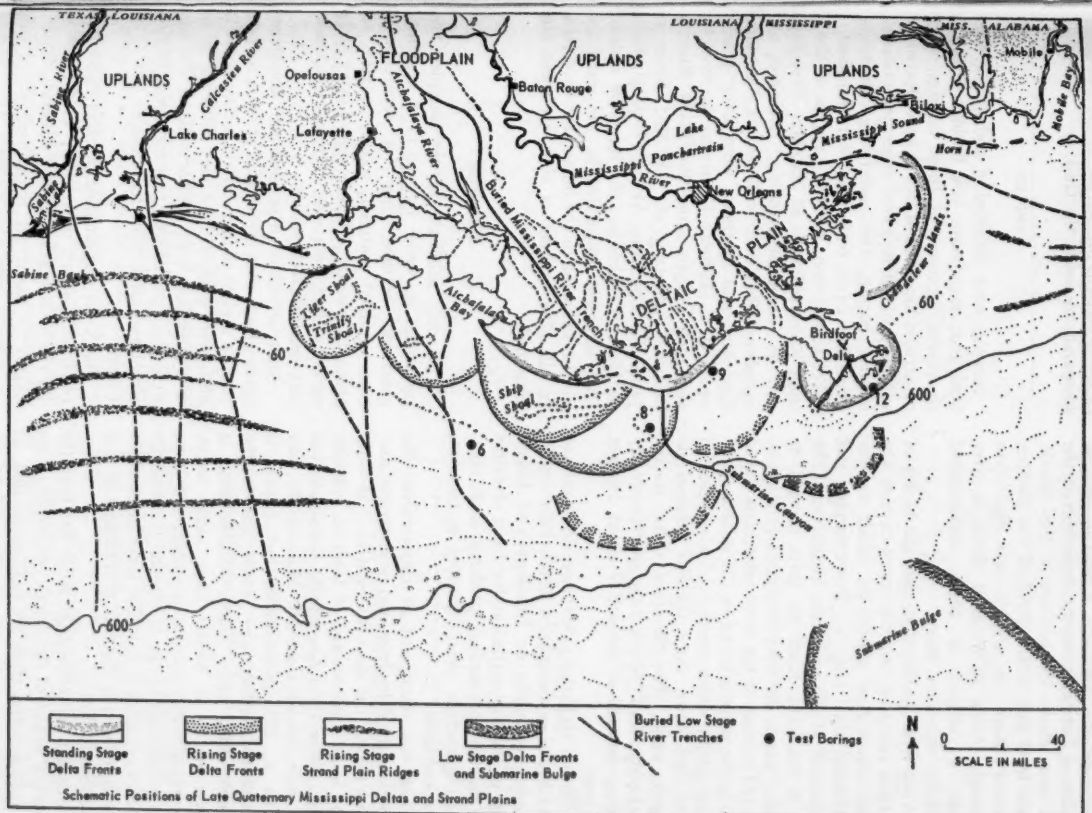


FIGURE 1.—FEATURES OF LATE QUATERNARY MISSISSIPPI RIVER DELTAIC DEPOSITION
After Fisk and McClelland, 1959

River flowed through a trench cut into a weathered oxidized Pleistocene surface to the Gulf of Mexico. Figure 1, a map illustrating features of Late Quaternary deposition in the shelf area, shows the position of this trench. As sea level rose with retreat of the glaciers, sediment filled the trench, and the river migrated. Owing to frequent shifts in river course, a complex of deltas accumulated along the regressive shore.

As sea level continued to rise, the mouth of the river retreated, and the water covered the deltas which now appear as shoals on the continental shelf (Fig. 1). Downwarping and compaction of the sediments accompanied the rise in sea level. After sea level reached its present stand, the river continued to migrate. The old courses of the river have been traced both in the sediments and on the surface of the deltaic plain, which consists of a complex of deltas laid down since the sea reached its present stand. The Birdfoot delta started developing about 450 years ago and is currently growing seaward at the rate of about 250 feet per year.

Fisk and McFarlan (1955) have distinguished a number of environmental facies of the deltaic complex, primarily on lithology and fauna. Most of the clay settled at the front of deltas and in the surrounding prodelta shelf area. Lesser quantities were deposited in marshes, bays, and lakes.

From the standpoint of clay-mineral distribution, the delta-front and prodelta environments appear the most informative. The delta front is an environment of fresh to brackish water where clays are rapidly deposited, not far from the river mouth. Most of the clay settles in this environment, but a part is transported farther and settles out in the prodelta shelf area. There, sedimentation is slower and is in deeper, more saline water. Accordingly, clay-mineral assemblages in these two environments can be studied for evidence of marine diagenesis as well as for changes which may be related to distance of transport or rate of sedimentation.

Although the Mississippi River has been the principal source of the deltaic sediments of the Louisiana coast, there are indications that, at particular times in the past, sediments from the Red River have predominated in localized areas. Mississippi River sediment is usually grayish, whereas clays carried by the Red River are characteristically reddish brown. Beds of reddish-brown clays of Late Quaternary age occur in borings in the Louisiana area. Apparently the Red River deposited these beds during periods in which this river, not commingling with the Mississippi River, maintained a separate course to the Gulf.

Samples

Samples were obtained of undisturbed cores from four foundation test borings. Figure 1 shows the locations of these borings, which penetrated hundreds of feet of Late Quaternary sediments of different environmental facies and, in one boring, the underlying Pleistocene formation. In all, 60 samples were examined.

For comparison, we took samples of Mississippi River sediment at Natchez, Mississippi, and at Tarbert's Landing, Louisiana, and samples of Red River sediment at Alexandria, Louisiana, and upstream from Lake Texoma where U. S. Highway 77 crosses the river. We also took samples from the Black River, a principal tributary of the Red River, at Jonesville, Louisiana. At each location, except for Tarbert's Landing, the sampling included both sediment suspended in the water and surface sediment in the immediate flood plain. In all, nine samples of river sediment were examined.

Methods

The clay minerals in the samples were suspended in water by use of a hypersonic generator. Settling time for separation of coarse-grained material, mostly quartz, was about 15 minutes. The purpose of using such a short settling time was to minimize possible clay-mineral fractionation.

Some of the suspension of fine-grained material was then evaporated at room temperature on a glass slide to obtain a clay film. The films were examined for clay minerals by use of a recording X-ray diffractometer. The films were also X-rayed after treatment with ethylene glycol and after heat treatment at 600° C. for 1 hour.

The X-ray examination of clays under these conditions is part of a widely used procedure for the identification of clay minerals. Identification is based on determination of basal spacings between silicate sheets. These spacings are calculated from the angular positions of X-ray diffraction maxima or "reflections," such as those shown in Figures 2–4. A later section of this report presents an interpretation of the patterns shown in these figures.

Montmorillonite, when air-dried, has a basal

spacing between 12 and 15 Å which expands to 17.7 Å with ethylene-glycol treatment and collapses to about 10 Å when heated to 600° C. Illite has a 10-Å spacing which is unaffected by

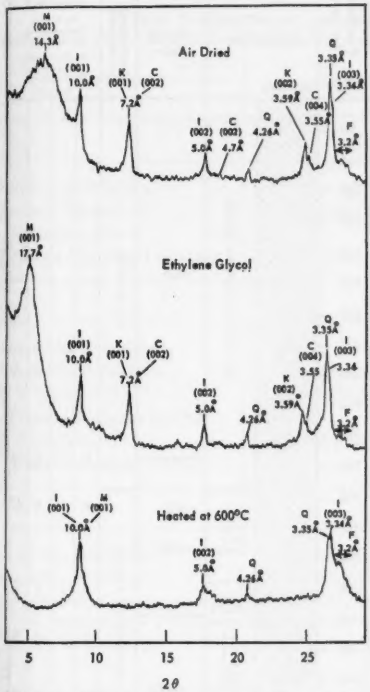


FIGURE 2.—X-RAY DIFFRACTION PATTERNS OF SAMPLE AT 246 FEET SUBSEA IN BORING NO. 12
CuKa radiation. C—chlorite; F—feldspars; I—illite; K—kaolinite; M—montmorillonite; Q—quartz

ethylene glycol or the heat treatment. Kaolinite's basal spacing at about 7 Å is unaffected by ethylene glycol but is destroyed by 600° C. heat. Chlorite is distinguished by a basal spacing of 14 Å which usually gives a relatively weak first-order reflection compared to higher-order reflections at 7 Å, 4.7 Å, and 3.5 Å. Chlorite does not expand with ethylene glycol, but heating at 600° C. usually enhances the first-order reflection and weakens the higher-order reflections. Randomly interstratified, or "mixed-layer," clays give basal reflections intermediate in position between those of the component clay minerals.

Results obtained by means of X-ray diffraction techniques are greatly limited because X-ray analyses of clays are at best semi-quantitative. The intensity of X-ray-diffraction maxima obtained with clay minerals varies with grain size, degree of orientation of clay

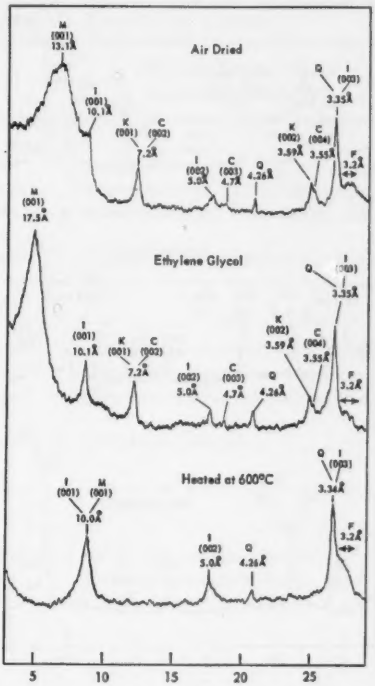


FIGURE 3.—X-RAY DIFFRACTION PATTERNS OF SAMPLE AT 149 FEET SUBSEA IN BORING NO. 6
CuKa radiation. C—chlorite; F—feldspars; I—illite; M—montmorillonite; K—kaolinite; Q—quartz

particles, and a number of other factors, none of which can be controlled readily or corrected for by techniques currently in use. For this reason, little significance can be attached to minor variations of reflection intensities.

In the present work, the relative abundance of different clay minerals was estimated from visual observation of reflection intensities, with allowance for diffraction characteristics of different clay minerals.

X-ray analysis of clays is further complicated by the difficulty of distinguishing chlorite and kaolinite in samples containing both minerals. The first- and second-order basal reflections of

kaolinite at about 7 Å and 3.5 Å, respectively, fall very close to or coincide with the positions of the second- and fourth-order and reflections of chlorite. The third-order chlorite reflection

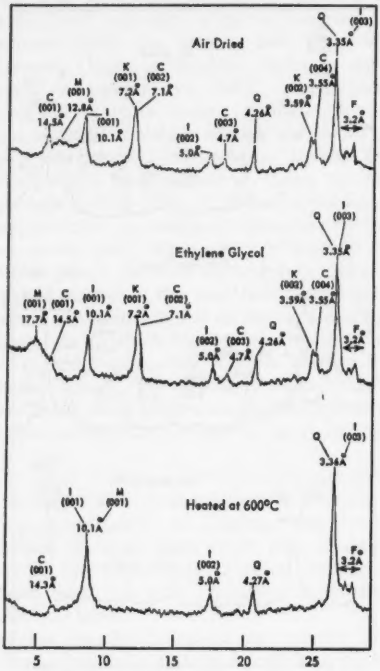
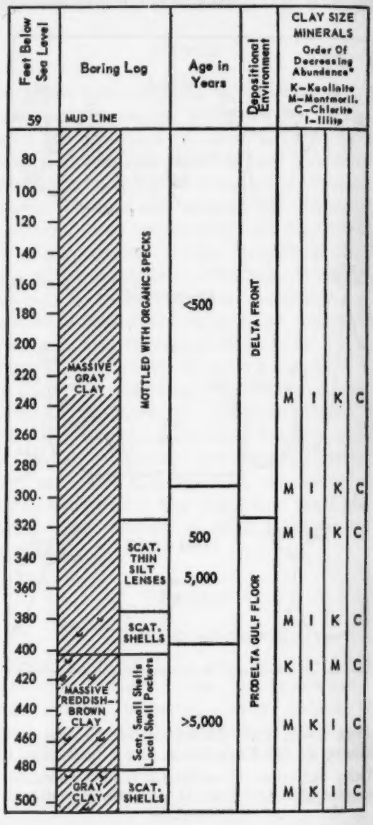


FIGURE 4.—X-RAY DIFFRACTION PATTERNS OF SAMPLE AT 419 FEET SUBSEA IN BORING NO. 12
CuKa radiation. C—chlorite; F—feldspars; I—illite; M—montmorillonite; Q—quartz; K—kaolinite

at about 4.7 Å is unique but may be weak. Heat treatment at 600° C. removes the kaolinite peaks but also weakens or removes the higher-order chlorite reflections. The identification of kaolinite, therefore, depends on fortuitous resolution of its reflections from those of chlorite or on the chlorite content being so small that the chlorite reflections do not interfere appreciably. The essential characteristic of chlorite is its reaction to heat, whereby the relative intensities of first- and higher-order basal reflections are altered. Consequently, resolution of higher-order reflections from reflections of kaolinite assists in the identification of chlorite also. The unique third-order reflection or the appearance with heating of a 14-Å peak, however, usually gives some indication of the presence of chlorite.



*All samples also contained quartz and feldspars

FIGURE 5.—DEPTH RELATIONSHIP OF CLAY-MINERAL ANALYSES TO LITHOLOGY, AGE, AND DEPOSITIONAL ENVIRONMENT, BORING NO. 12

RESULTS

Identities of Clay Minerals in Samples

All the deltaic samples examined contained montmorillonite, kaolinite, illite, and chlorite. The river-borne solids contained the same clay minerals, although all four were not detected in two of these samples.

The identification of montmorillonite and of illite in the samples by X-ray diffraction techniques presents no difficulties. As mentioned

in a preceding section, however, the identification of kaolinite and chlorite, when both are present in a single sample, requires an interpretation of the X-ray data which sometimes involves complexities. Figures 2, 3, and 4, which represent the types of X-ray diffraction patterns obtained from both deltaic and river samples, illustrate the interpretation procedure.

Figure 4 illustrates patterns which clearly indicate the presence of both chlorite and kaolinite. The first-, third-, and fourth-order peaks of chlorite at about 14.5 Å, 4.7 Å, and 3.5 Å, respectively, are discernible before heating, and the first-order peak is discernible after heating. In the case of kaolinite, the second-order peak has been resolved. Further, the reflection at about 7 Å is too large to be due entirely to chlorite, which from the other reflections appears to be less concentrated. Figures 2 and 3 illustrate patterns that show fewer indications of chlorite. The third-order peaks are less distinct in these patterns, and no 14-Å peaks developed with heating. Most of the deltaic samples gave patterns similar to those of Figures 2 and 3.

Although the literature suggests that mixed-layer clay minerals are common constituents of sediments, only 3 of the 69 samples examined gave fairly definite indication of such materials. Two of the samples from deltaic sediments and one from the Black River may have contained small amounts of a montmorillonite-illite clay The X-ray diffraction patterns of these samples showed small peaks at about 14Å after ethylene-glycol treatment which disappeared after heat treatment at 600° C. No other peaks indicative of mixed-layer clays were observed.

TABLE 1.—ABUNDANCE OF CLAY MINERALS IN RIVER-BORNE SAMPLES

Location	Sample type	Clay minerals in order of decreasing abundance*
Mississippi River		
Natchez, Mississippi	Water-borne sediment	montmorillonite, kaolin, illite, chlorite
Natchez, Mississippi (Louisiana side)	Flood-plain sediment	montmorillonite,† illite, kaolin, chlorite
Tarbert's Landing, Louisiana	Water-borne sediment	montmorillonite, kaolin, illite, chlorite
Red River		
Alexandria, Louisiana	Water-borne sediment	montmorillonite, kaolin, illite, chlorite
Alexandria, Louisiana	Flood-plain sediment	illite, kaolin, chlorite, montmorillonite
U. S. Highway 77 (Texas-Oklahoma border)	Water-borne sediment	illite, kaolin, montmorillonite, chlorite
U. S. Highway 77 (Texas-Oklahoma border)	Flood-plain sediment	illite, kaolin, chlorite
Black River		
Jonesville, Louisiana	Water-borne sediment	kaolin, illite, montmorillonite
Jonesville, Louisiana	Flood-plain sediment	illite, kaolin, montmorillonite, chlorite

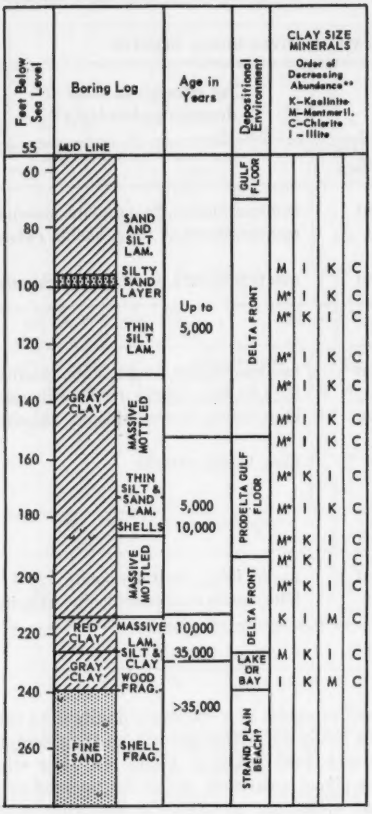
*all samples also contained quartz and feldspars
† very abundant

Abundances of Clay Minerals

Figures 5–8 give the results of X-ray analyses of deltaic samples and show also the depositional environment, lithology, and approximate age of sediments penetrated in each of the borings. In these figures, the clay minerals are listed in the order of decreasing relative abundance.

On the basis of relative abundance, variation in montmorillonite content was the most outstanding and definite difference in these samples. A preponderant number, 47 of the 60 deltaic samples, had montmorillonite as the most abundant clay mineral. In these samples, illite and kaolinite followed mont-

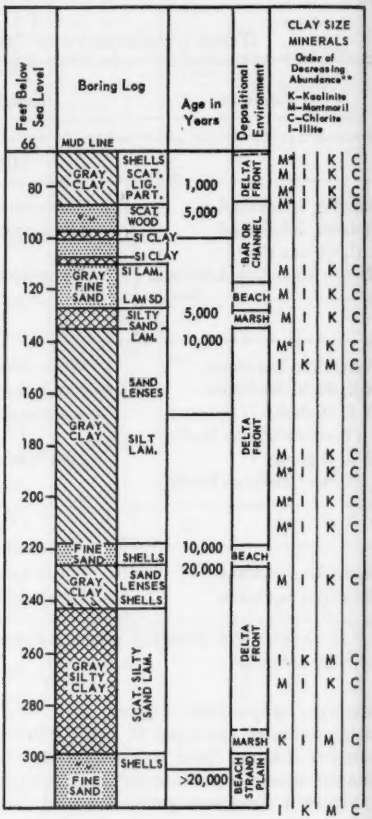
morillonite in abundance, with illite somewhat more abundant in most of the samples. Chlorite was the least abundant of the clay minerals.



* Very Abundant
** All samples also contained quartz and feldspars

FIGURE 6.—DEPTH RELATIONSHIP OF CLAY-MINERAL ANALYSES TO LITHOLOGY, AGE, AND DEPOSITIONAL ENVIRONMENT, BORING NO. 9

sediment. The X-ray diffraction patterns of Figure 4 were obtained from a sample representative of the low-montmorillonite type of



* Very Abundant
** All samples also contained quartz and feldspars

FIGURE 7.—DEPTH RELATIONSHIP OF CLAY-MINERAL ANALYSES TO LITHOLOGY, AGE, AND DEPOSITIONAL ENVIRONMENT, BORING NO. 8

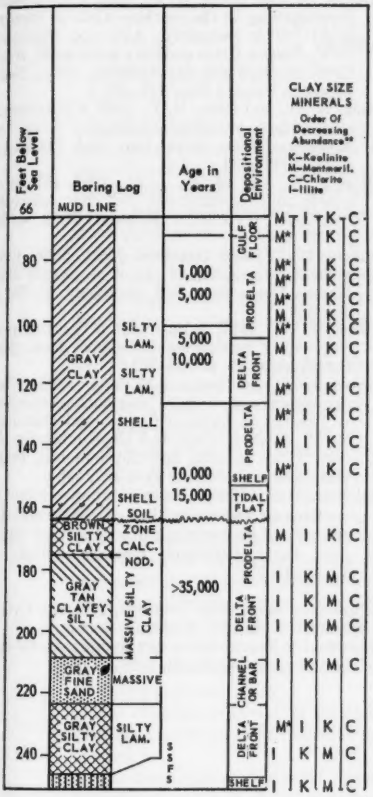
Figures 2 and 3 illustrate the X-ray diffraction patterns obtained with this type of clay. All the samples in this category, except one from the oxidized Pleistocene surface penetrated in boring no. 6, were gray.

Montmorillonite was not the most abundant clay mineral in the remaining deltaic samples. The abundances of clay minerals other than montmorillonite, however, were comparable with those of the high-montmorillonite type of sediment. Of the 13 samples which were of the low-montmorillonite type, 8 were grayish and 5 were reddish.

Table 1 shows the relative abundances of clay minerals in samples from the Mississippi, Red, and Black rivers and from adjacent flood plains.

Montmorillonite was the most abundant clay mineral in the Mississippi River samples,

which were uniformly gray. This mineral was the most abundant in only one of the four Red River samples, however. All the latter samples had a characteristic red color, and, except for the one sample noted, contained either a relatively small amount of montmorillonite or else none at all. The samples from the Black River were also relatively low in montmorillonite content. Generally, then, Mississippi River sediments were of the high-montmorillonite type, and Red River and Black River solids were of the low-montmorillonite type.



* Very Abundant
** All samples also contained quartz and feldspars

FIGURE 8.—DEPTH RELATIONSHIP OF CLAY-MINERAL ANALYSES TO LITHOLOGY, AGE, AND DEPOSITIONAL ENVIRONMENT, BORING NO. 6

INTERPRETATION OF RESULTS

Evidence on Marine Diagenesis and Fractionation by Sedimentation

If, in a marine environment, montmorillonite alters to form illite or chlorite, it might be expected that marine sediments would contain appreciably less montmorillonite than their nonmarine equivalents. The results of this study show no differences in montmorillonite abundance attributable to diagenesis.

For example, in boring no. 9 (Fig. 6), samples from two delta-front facies, one prodelta Gulf floor facies, and a lake or bay facies, were high in montmorillonite with two exceptions. Of these two, one was a sample of a delta-front facies and the other a sample of a lake or bay facies.

The patterns of Figures 2 and 3 illustrate, in detail, the similarity of the high-montmorillonite type of clay in sediments deposited in both marine and nonmarine waters. The patterns of Figure 2 were obtained with a sample of the sediment deposited rapidly at the front of the modern Birdfoot delta in fresh to brackish water. On the other hand, the patterns of Figure 3 were obtained with a sample of sediments deposited slowly in the marine water of the shelf area. Even though the samples represented environmental extremes, the two sets of patterns are nearly identical.

Accordingly, it is concluded that either the clay minerals introduced into this area of the Gulf have not altered as a result of deposition in a marine environment or that any alteration has been too slight to permit detection by X-ray-diffraction techniques.

Similarly, examination of the data shows no differences in clay-mineral abundances which can be related to fractionation of clay minerals by sedimentation. Assemblages and abundances of clays found in sediments relatively far removed from the delta were entirely similar to those of sediments deposited rapidly in the immediate vicinity of the river mouth.

Influence of Source

Since the differences observed in clay-mineral abundances in the deltaic samples cannot be related to diagenetic changes or to fractionation by sedimentation, they may result from differences in source of the sediments brought into the delta region.

The geologic evidence shows that most of

the clays deposited in the delta complex have been transported by the Mississippi River. The results of this investigation are in accord, in that most of the deltaic sediments examined were high in montmorillonite and were gray, as are the sediments transported by the Mississippi River. The geologic evidence also shows that at various times in the past, sediments from the Red River have predominated in localized areas of the delta region. Paralleling this geologic evidence, all the reddish-brown sediments from the borings in the delta (except for one sample from the weathered Pleistocene surface) which were examined were of the low-montmorillonite type, similar to Red River sediments.

Although all samples of the post-Pleistocene deltaic sediments which were reddish were of the low-montmorillonite type, not all the low-montmorillonite samples were reddish. Possibly the gray low-montmorillonite samples also reflect source; there is not sufficient information available to substantiate this point.

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STRATIGRAPHIC AND GEOTECTONIC RELATIONSHIPS IN NORTHERN VERMONT AND SOUTHERN QUEBEC

BY WALLACE M. CADY

ABSTRACT

Stratified rocks of early and middle Paleozoic age form a belt of northeast-trending anticlinoria and synclinoria of middle Paleozoic age, in northern Vermont and adjacent parts of southern Quebec. The foreland margin of this belt, in the Champlain and St. Lawrence valleys to the west, is cut by eastward-dipping thrust faults of middle Paleozoic age and by later northeast-trending normal faults. The Green Mountain anticlinorium, which is the dominant structure of the region, is flanked to the west, on the foreland, by the St. Albans-Hinesburg-Middlebury synclinorium and to the east, in the midst of the folded belt, by the Connecticut Valley-Gaspe synclinorium. The principal thrust faults, notably the Champlain and Philipsburg thrusts, are in the west limb of the St. Albans-Hinesburg-Middlebury synclinorium. East of the Connecticut Valley-Gaspe synclinorium is the Boundary Mountain anticlinorium, in eastern Vermont and adjacent New Hampshire and along the international boundary between Quebec and Maine.

Two contrasting intergradational lithic assemblages, the graywacke-shale assemblage and the carbonate-quartzite assemblage, characterize the protolith of the bedded rocks. The graywacke-shale assemblage includes thick sections of lower Paleozoic strata, portions of which lap both gradationally and unconformably westward on the foreland, particularly in Quebec; it also includes middle Paleozoic strata that offlap eastward away from the foreland. The carbonate-quartzite assemblage laps both unconformably and gradationally eastward over the graywacke-shale assemblage in sections of the middle Paleozoic east of the axis of the Green Mountain anticlinorium.

Stratigraphic correlation has become well established in the foreland belt where numerous distinctive and fossiliferous strata, chiefly of the carbonate-quartzite assemblage, have escaped metamorphism. It is also fairly clear in sections in the eastern foreland and western part of the Green Mountain anticlinorium, where the strata of the carbonate-quartzite assemblage extend eastward and interfinger with rocks of the graywacke-shale terrane. Rocks that are entirely of the graywacke-shale assemblage have been correlated in the present study.

The stratified rocks west of the axis of the Green Mountain anticlinorium are of Cambrian(?), Cambrian, and Ordovician age; those to the east range in age from Cambrian to Devonian.

The geotectonic setting of the region is the once mobile belt of the Appalachian orthogeosyncline, which is at the southeastern margin of the stable continental block, or craton, of North America. The orthogeosyncline was a belt, chiefly of subsidence, that embraced two parallel and adjoining longitudinal zones: the eugeosynclinal zone, which was more mobile, and the miogeosynclinal zone, which was less mobile. Second- and third-generation geosynclines are superimposed not only on the orthogeosyncline but also on adjoining parts of the craton.

Uplift, and finally folding, gradually superseded subsidence in the orthogeosyncline. Local uplift, chiefly within the eugeosynclinal zone, provided most of the clastic sediments, principally those of the graywacke-shale assemblage. General stabilization of the western part of the orthogeosyncline at the end of the Ordovician was accompanied by eastward migration of the miogeosynclinal zone. Localized uplift within the eastern part of the orthogeosyncline at this time is marked by unconformities referred to the Taconic disturbance. Folding and uplift after the Early Devonian is shown by angular unconformities referred to the Acadian and Appalachian orogenies.

The interpretation of the geotectonic relations of the bedded rocks is aided by critical features of the magmatic activity that began with, accompanied, and followed the diastrophism.

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INTRODUCTION

Purpose

The purpose of the present study has been to clarify the stratigraphic relationships in Vermont and between Vermont and Quebec and to provide a better comprehension of the geotectonic setting of the stratified rocks of northwestern New England and adjacent areas in southeastern Quebec (Fig. 1). This study has been made in connection with a regional investigation of the ultramafic intrusive rocks and related talc and asbestos deposits in north-central Vermont.

Related Work

The pioneer geologists in the northern New England states and Quebec extended major stratigraphic units along strike across the international boundary and correlated rocks of contrasting lithology east-west across the strike. Most notable of these earlier workers were Logan in Quebec (Logan and Hall, 1865) and Hitchcock (1877) in New England. Others, notably Ells in Quebec (1887; 1888; 1896) and Richardson in Vermont (1908), have attempted correlations across the strike that are based chiefly on lithologic similarities. In the course of more detailed work, smaller stratigraphic units were identified in widely separated areas, and differences between units became more apparent than similarities. These smaller units, many of which were first described far to the north in Quebec or to the south in Vermont, failed to match near the boundary. This fact, as well as the diverse facies of the smaller units, has tended to obscure the relations between the geology of Vermont and that of Quebec.

The work of Booth (1950) to the west of the Green Mountains and of Doll (1951) to the east of the mountains led to the first modern reports on the geology of northern Vermont. These reports provided a basis for detailed correlations with the geologic features of southern Quebec. The investigation by the U. S. Geological Survey that is now in progress consists of geologic mapping of an extensive

belt of metamorphosed sedimentary, volcanic, and ultramafic intrusive rocks in the Green Mountains of central and northern Vermont.

Scope and Method of the Present Study

The present study has been planned as a synthesis of various field investigations of many geologists, both published and unpublished. The author has tried to include in the correlation table (Pl. 3) as much as possible of the basic information on lithology, correlation, and age.

Inasmuch as geologic maps that show the distribution of formations only partly cover the region the author presents simply a generalized tectonic map (Pl. 1), in which the rocks of various geologic ages are differentiated. For further details the reader is referred to maps in the sources cited at the bottom of the correlation table (Pl. 3) and also in the text. Various localities outside the area of the tectonic map are shown on the index map of southeastern Quebec and northwestern New England (Fig. 1).

The quality and quantity of information cited in this study vary primarily with the accessibility of the rocks. Much of the area in Vermont east of the Champlain Valley is in steep and heavily wooded mountains where detailed geologic investigations have been delayed until very recent years. Data from the Green Mountains in Vermont and from their northern extension, the Sutton Mountains in nearby Quebec, are just beginning to be published. The author has attempted to make up for this deficiency as much as possible by reconnaissance studies in areas that seem most likely to yield critical information. Investigations of radiometric age, not only of the stratigraphic units but of their metamorphism, which are just beginning, will be a significant part of any definitive discussion of the region.

Acknowledgments

Extensive acknowledgments are in order inasmuch as this paper is largely the outcome of consultations with many geologists both in the field and during preparation of the manuscript.

The author wishes first to acknowledge the part that P. H. Osberg, of the University of Maine, and the author's associates, A. L. Albee and A. H. Chidester, of the U. S. Geological Survey, have taken in this study. Each has participated with the author in related field and laboratory investigations and has contributed extensively to the facts and interpretations presented here. Osberg, Chidester, Albee, W. S. White, and F. F. Osborne reviewed successive drafts of the manuscript, and M. P. Billings and Marshall Kay acted as critics for The Geological Society of America. The author is indebted to all these men for their corrections of fact and suggestions for improved presentation.

Profs. M. P. Billings and J. B. Thompson, Jr., of Harvard University, and their associates and students have contributed much to the interpretation of the metamorphic stratigraphy and structural geology of the region. Prof. F. F. Osborne of Laval University and his students have provided the author with copies of thesis manuscripts that are valuable supplements to the published geologic literature on the Eastern Townships of Quebec.

Field geologists of the Quebec Department of Mines have helped in many ways. Our knowledge of the Canadian extensions of the rocks in Vermont was much enhanced by the late H. C. Cooke, who for several years led the geologists of the U. S. Geological Survey on annual excursions in the Eastern Townships. Professor Osborne and his associates of the Quebec Department of Mines, J. R. Béland, F. W. Benoit, Henry De Romer, Gilles Duquette, W. A. Gorman, P. E. Grenier, P. J. Lesperance, R. A. Marleau, and C. W. Stearn, conducted several instructive field trips. The author is particularly indebted to Béland for time given in the Matapedia and Gaspe regions.

Members of the staff of the Vermont Geological Survey have conducted numerous field conferences in which the author and his associates have participated; these have furnished extensive information on the geology of northern Vermont. The help and suggestions of C. G. Doll, State Geologist, and of S. P. Averill, R. A. Christman, J. G. Dennis, E. H. Ern, B. K. Goodwin, L. M. Hall, R. H. Konig, V. R. Murthy, and D. T. Secor are herewith acknowledged.

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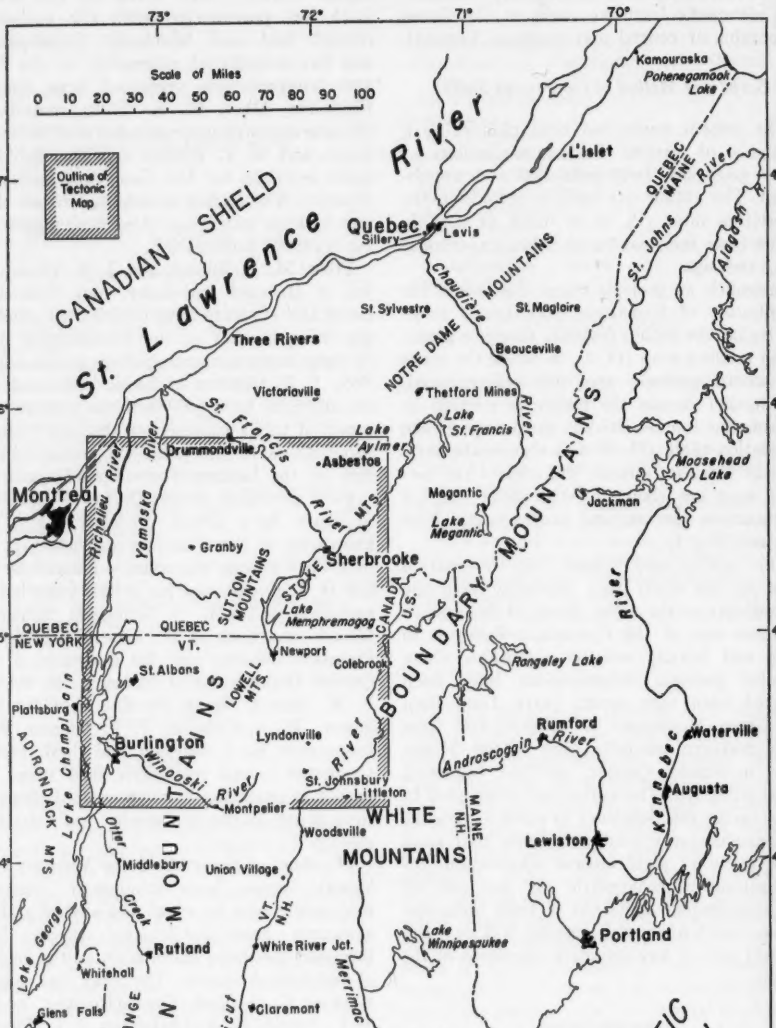

FIGURE 1.—INDEX MAP OF SOUTHEASTERN QUEBEC AND NORTHWESTERN NEW ENGLAND

Lyons, J. A. MacFadyen, Jr., P. B. Myers, M. J. Rickard, P. H. Riordon, John Rodgers, J. L. Rosenfeld, A. B. Shaw, J. W. Skehan, S. W. Stone, George Theokritoff, C. W. Welby, and E-an Zen.

General Setting

The stratified rocks, which are of Paleozoic age, are in the northern part of the Appalachian folded belt. Anticlinoria and synclinoria trend northeastward within the folded belt. The foreland, which is west of the belt of principal folding but includes the western most of the synclinoria, is characterized by eastward-dipping thrust faults. Both the folds and the thrust faults are of Paleozoic age. The rocks of the foreland adjoin and overlap westward upon the Precambrian terranes of the Adirondack Mountains and of the Canadian Shield, in the Champlain and St. Lawrence valleys. Normal faults, which trend northeast, intersect the western foreland and the Precambrian terranes; these faults are of Paleozoic age or younger.

The Green Mountain[1] anticlinorium (Pl. 1) is the dominant structural feature of the region (Adams, 1846, p. 167–168; Cady, 1945, p. 564; Chidester, 1953; Hitchcock and others, 1861, p. 254, 464–467). It extends more than 300 miles from the southern boundary of Vermont north-northeast through the State and into Quebec (Cooke, 1954; Selwyn, 1879, p. 7–9). (*See* Fig. 1.) Various moderately persistent large anticlines, in which Lower Cambrian rocks are exposed, are referred to here as the axial anticlines of the anticlinorium. Some of these anticlines are placed en echelon rather than linearly, and most of them coincide with the crest line of the Green Mountains. Several lesser tectonic units in addition to the axial anticlines are included in the anticlinorium in northern Vermont and nearby Quebec (Pl. 1).

The axial anticline of the Green Mountain anticlinorium in the vicinity of the international boundary is flanked on the west by a very narrow synclinal tract, which expands into a wider syncline both to the north and south of the boundary. This feature is here referred to as the Cambridge syncline for its occurrence in Cambridge township, Vermont, about 25 miles south of the boundary. The Enosburg Falls anticline, which is named for a village in northern Vermont located near its axis, is west of the Cambridge syncline but also in the Green Mountain anticlinorium.[2] Subsidiary anticlines and synclines also occur east of the axial anticline of the Green Mountain anticlinorium. They include synclines near Lake Memphremagog in Quebec (Logan, 1863, p. 251, 435–436) and fairly persistent anticlines in the Lowell Mountains in Vermont and in the Stoke Mountains northeast of Lake Memphremagog (Dresser, 1913, p. 52; Ells, 1887, p. 31–32, map; 1896, map).

Flanking the Green Mountain anticlinorium are major synclinoria. The St. Albans-Hinesburg-Middlebury synclinorium lies to the west, on the eastern margin of the foreland, and the Connecticut Valley-Gaspe synclinorium is to the east, in the midst of the folded belt.

The St. Albans-Hinesburg-Middlebury synclinorium is segmented transversely into three parts by culminations. The three segments have been named individually the St. Albans,[3] Hinesburg, and Middlebury synclinoria (Cady, 1945, p. 562–564; Shaw, 1958, p. 563–565). The Middlebury synclinorium is south of the area of the tectonic map (Pl. 1).

East of the St. Albans and Hinesburg synclinoria, in the west limb of the Enosburg Falls anticline, is a zone of eastward-dipping thrust faults that has been variously mapped and interpreted (Booth, 1950, Pl. 1, p. 1160–1166; Cady, 1945, Pl. 10, p. 566–567; Clark, 1934, Fig. 2, p. 9; Shaw, 1958, Pl. 1, p. 557–559; S. W. Stone, 1951, Ph.D. thesis, Harvard Univ., Pl. 17). The faults do not now appear to be as extensive or as numerous as originally believed and possibly disappear in Quebec a few miles north of the international boundary.[4]

[1] The designation, Green Mountain-Sutton Mountain anticlinorium, which is also common and recognizes the northward extension of the anticlinorium into Quebec, is not used here because the repeated syllables have proved monotonous to read.

[2] Field studies completed by R. A. Christman, D. T. Secor, and the author in the fall of 1959 show that the Enosburg Falls anticline near the Winooski River is as much as 3 miles west of the position inferred on Plate 1. The Cambridge syncline in this vicinity is split into two synclinal tracts by a local anticline. This relationship was found after Plate 1 was prepared.

[3] The St. Albans synclinorium includes the "Rosenberg slice" of Clark (1934, p. 4, 8) and the northern part of the "central sequence" of Keith (1932, p. 357, Fig. 1, p. 369–379).

[4] Field studies, by J. G. Dennis north of the Winooski River and by the author to the south, were completed in the fall of 1959 after the tectonic map (Pl. 1) was prepared.

West of the St. Albans-Hinesburg-Middlebury synclinorium are the Champlain and Philipsburg thrusts, which also dip east. The extensive Champlain thrust dies out a little north of the international boundary, but near the boundary it separates the synclinorium from various foreland thrust slices to the west. These slices are in turn separated by eastward-dipping thrust faults (Clark, 1934, p. 6–8; 1951, p. 19–20; McGerrigle, 1930, p. 181–186).

The foreland thrust slices include the Philipsburg slice and the Highgate Springs-St. Dominique slice, in the vicinity of the international boundary. The Philipsburg slice contains a homoclinal section that dips eastward and is based on the Philipsburg thrust. The Highgate Springs-St. Dominique slice contains a folded section that is based on the Highgate Springs-St. Dominique thrust; it is bounded to the east by the Philipsburg slice at the international boundary and by the Champlain thrust, farther south.

The Champlain-Richelieu foreland is a belt of the foreland that is west of the synclinoria and major thrust faults (Hawley, 1957, Pl. 1). The section in this part of the foreland is characterized by many small folds. It is cut by normal faults that strike north-northeast and by minor thrust faults that dip east.

The Connecticut Valley-Gaspe synclinorium (Pl. 1) occupies a broad and extensive tract that spreads eastward across the upper Connecticut River valley into New Hampshire and from Long Island Sound north and northeast into the Gaspe Peninsula, Quebec (McGerrigle, 1950, p. 105; (John Rodgers, 1957, oral communication). This tract was first distinguished by Logan and Hunt (1855, p. 52–53) and was presumed to be grossly synclinal (Ells, 1887, p. 24–25, map). The axial region of the Connecticut Valley-Gaspe synclinorium in Vermont includes the Willoughby arch, an anticlinal structure that is related to doming brought about by intrusion of granitic plutons after the climax of the folding that produced the synclinorium (Dennis, 1956, p. 67–69; Eric and Dennis, 1958, p. 53–54). (*See* Lyons, 1955, p. 125; White and Jahns, 1950, p. 214–219.) Numerous granitic plutons (not shown on Plate 1) that crop out in the Connecticut Valley-Gaspe synclinorium in northeastern Vermont further substantiate such an interpretation.

The Boundary Mountain anticlinorium forms the eastern limit of the Connecticut Valley-Gaspe synclinorium in eastern Vermont and adjacent New Hampshire (Pl. 1) and along the international boundary between Quebec and Maine (Fig. 1) (Dresser, 1913, p. 52; Marleau, 1958, p. 3). The Gardner Mountain anticline (Pl. 1) is the axial anticline of this anticlinorium southeast of St. Johnsbury, Vermont (Billings, 1937, p. 519–520; Eric and Dennis, 1958, p. 29–30).

Lithic Assemblages and Facies Relationships

General Statement

Two contrasting intergradational lithic assemblages characterize the bedded rocks of the Vermont-Quebec region. For simple reference they are here designated the graywacke-shale assemblage and the carbonate-quartzite assemblage. These lithic assemblages are differentiated on Plates 2 and 3. Plate 3, the correlation table, has a double column for each structural belt; sections in Quebec are in the left column, and sections south of the boundary are in the right column.

The graywacke-shale assemblage is made up of varied facies of graywacke, shale, chert, and some quartz sandstone and arkose. It includes a continuous section of Cambrian(?), Cambrian, and Ordovician strata probably as much as 10 miles thick in the Green Mountain anticlinorium of Vermont. This assemblage laps both gradationally and unconformably westward on the foreland, particularly in Quebec. Near the international boundary more than 3000 feet of strata that are chiefly of Ordovician age and of the graywacke-shale assemblage overlap westward unconformably on the carbonate-quartzite assemblage of the foreland. Silurian and Devonian rocks of the graywacke-shale assemblage offlap eastward away from the foreland.

Mafic to felsic volcanic rocks are interbedded with the graywacke-shale assemblage, principally in and east of the Green Mountain anticlinorium, where they are possibly as much as 5 per cent of the rock. The mafic volcanic rocks are composed of the weathered and eroded detritus of lava flows as well as flows intact.

The carbonate-quartzite assemblage includes limestone, shaly limestone, calcareous shale, dolomite, quartz sandstone (orthoquartzite), and their conglomeratic facies. This assemblage is typically developed in sections of Cambrian and Ordovician strata as much

as 8000 feet thick, west of the Green Mountain anticlinorium in Vermont and nearby parts of Quebec. The carbonate-quartzite assemblage also occurs in sections of Silurian and Devonian strata east of the axis of the Green Mountain anticlinorium; there it is as much as 6000 feet thick and laps both unconformably and gradationally eastward over the graywacke-shale assemblage.

The rocks of the graywacke-shale assemblage are in general more altered by metamorphism, and their protoliths therefore are less apparent than those of the carbonate-quartzite assemblage. This difference is partly due to original composition and partly to the occurrence of the graywacke-shale assemblage chiefly in and east of the Green Mountain anticlinorium, where the rocks are in thicker sections and were once more deeply buried, underwent greater deformation, and have been intruded by extensive plutons of igneous rocks. Schist, phyllite, slate, quartzite, granulite, gneiss, greenstone, and amphibolite are the common metamorphic derivatives of the graywacke-shale assemblage and associated volcanic rocks.

Plate 3 shows that the contrasting lithic assemblages—graywacke-shale to the east and carbonate-quartzite to the west—interfinger with each other in the Cambrian(?) and Cambrian sections of the eastern foreland and the western part of the Green Mountain anticlinorium, near the international boundary. Interrelationships of the contrasting rock assemblages are less apparent in various other structural belts, partly because critical stratigraphic units were eroded from anticlinorial tracts during later geologic history or remain at depth beneath synclinorial tracts.

Carbonate-Quartzite Assemblage of the Foreland

The rocks of the foreland (Cambrian and Ordovician) are chiefly of the carbonate-quartzite assemblage, which totals about 8000 feet thick in west-central Vermont (Cady, 1945, p. 524–525). Near the international boundary the total thickness of the Gilman quartzite, Dunham dolomite, Saxe Brook dolomite, Gorge formation, and Highgate formation is about the same as that of the assemblage in west-central Vermont (Shaw, 1958, p. 524, 527, 544, 549, 551).

The carbonate rock of Cambrian age is nearly all dolomite. Limestone is interbedded with the dolomite of Ordovician age and is most abundant in the eastern part of the foreland belt, where it is represented chiefly by calcite marble. Typical of the westward facies variation from limestone to dolomite is the change from the limestone of the St. Armand and Corey of the Philipsburg slice, through the interbedded dolomite and limestone of the Beldens formation in the Highgate Springs-St. Dominique slice, into the Bridport dolomite of the Champlain-Richelieu foreland.

Quartzite is chiefly of Cambrian age and includes several units, two of which are known to thicken westward in the foreland of west-central Vermont (Cady, 1945, p. 531, 536). Individual beds of quartzite are thicker and more numerous toward the west and grade laterally into and interfinger with dominantly dolomitic strata to the east. The Gilman quartzite is probably at depth in the sections on the foreland near the international boundary in northwestern Vermont and nearby Quebec. The direction in which the Gilman thickens is not known.

The carbonate-quartzite assemblage extends 1000–2000 feet higher in the section in the Champlain-Richelieu foreland than in the foreland section to the east. In the Champlain-Richelieu foreland it includes the Lacolle conglomerate, which is interpreted by Clark and McGerrigle (1936) as a basal conglomerate,[5] and interbedded argillaceous limestone and calcareous shale of the Cumberland Head and Stony Point formations. The Cumberland Head and Stony Point are probably western facies equivalents of the Hortonville slate in the Highgate Springs-St. Dominique slice and the Morses Line slate in the St. Albans synclinorium, though the facies relationships cannot be demonstrated because faults intervene.

Graywacke-Shale Assemblage of the Foreland

The uppermost rocks of the foreland (Middle and Upper Ordovician) are of the graywacke-shale assemblage. They occur chiefly in the western part of the foreland (Champlain-Richelieu), having been eroded from much of the eastern foreland, particularly in Vermont. The thickness in the northern Champlain Valley in Vermont (Middle Ordovician) ranges from 1000 to 2100 feet (Hawley, 1957, p. 58); the upper rocks in the section have been eroded. The upper rocks (Middle and Upper

[5] Stone (1957) called this conglomerate a fault breccia.

Ordovician) are, however, exposed to the north in Quebec, where they range from about 3000 to about 5000 feet thick (Clark, 1947, p. 10–11; 1955, p. 16, 20).

The Iberville formation, a rhythmically interbedded clay shale and graded dolomitic siltstone, and the Hathaway formation, composed of argillite and bedded radiolarian chert with blocky inclusions of limestone, dolomite, dolomitic quartz siltstone, sandstone, coarse graywacke, and chert (Hawley, 1957, p. 55), make up the section in Vermont. The section in Quebec includes the following in addition to the Iberville formation: the Nicolet River formation, which is composed of gray shales with thin interbeds of limestone and sandstone; the Pontgravé River formation, composed chiefly of dark-gray shales; and the Bécancour River formation, made up mostly of red shale (Clark, 1955, p. 21, 33, 36). Rocks referred to the Lorraine group in the Highgate Springs-St. Dominique and Philipsburg slices possibly include eastern extensions of the lower of these formations, though most probably have been eroded in the eastern area.

Mixed Lithic Assemblages of the St. Albans Synclinorium and the Enosburg Falls Anticline

The graywacke-shale and carbonate-quartzite assemblages interfinger in northwestern Vermont and nearby Quebec. The total thickness (Cambrian? through Middle Ordovician) is about 10,000 feet, or a little more than that of the carbonate-quartzite assemblage in adjacent parts of the foreland (Booth, 1950, p. 1136; Shaw, 1958, p. 523). The thickness of the lower part of the section (Cambrian? and Lower and Middle Cambrian) is about one-tenth normal at some localities on the east limb of the Enosburg Falls anticline in Quebec, but nearly all formational units are present there (Clark, 1934, p. 10, Table 2).

The Tibbit Hill volcanics (Cambrian?) are near the base of the section in the Enosburg Falls anticline. The rock is principally greenstone, the protolith of which was probably either basalt or andesite. Pillow lavas occur in the Tibbit Hill volcanics 3.5 miles S. 63° W. of the village of Bakersfield, Vermont. Metarhyolite is interbedded with the greenstone 12 miles east of Granby, Quebec. Detrital facies of the Tibbit Hill grade upward and laterally into the Pinnacle formation where the Call Mill slate is absent.

The Call Mill slate, Pinnacle formation (graywacke, phyllite, and arkose), West Sutton slate, Oak Hill and Parker slates, St. Albans slate, Skeels Corners slate, and Hungerford slate are interstratified with the White Brook dolomite, Dunham dolomite, Rugg Brook dolomite (and sandstone), and Scottsmore quartzite, the Gorge formation, and the Highgate formation. The Gilman quartzite (Lower Cambrian), which is also part of this alternating sequence, is a "dirty" quartzite that is intermediate in type between graywacke and orthoquartzite and becomes phyllitic toward the east. Carbonate-quartzite units tend to thin to the east and disappear, whereas graywacke-shale units disappear westward (Pl. 2). At the top of the section is the Morses Line-Stanbridge slate (Middle Ordovician), which extends westward all the way across the St. Albans synclinorium; it is more calcareous than the underlying slates.

Conglomerates that contain phenoclasts chiefly of limestone and dolomite set in a matrix of limestone or dolomite are distinctive features. The Rugg Brook dolomite is partly conglomeratic; it contains boulders and matrix of sandy dolomite. The Mill River and Rockledge conglomerates are composed of phenoclasts of limestone set in a sandy limestone or dolomite matrix. The Gorge formation contains conglomerate and breccia with phenoclasts of both limestone and dolomite. The Mystic conglomerate (=?"Corliss") is composed of limestone phenoclasts in a fairly pure limestone matrix.

The section from the Parker slate up to the Rockledge conglomerate grades laterally southward into carbonate rocks and quartzite in the Hinesburg and Middlebury synclinoria of west-central Vermont (Cady, 1945, p. 534, 537–538). Units above the Rockledge conglomerate have been eroded from the culmination between the St. Albans and Hinesburg synclinoria, but presumably they, too, originally passed southward into rocks chiefly of the carbonate-quartzite assemblage.

Unconformities beneath the conglomeratic units—Rugg Brook formation, Mill River conglomerate and Rockledge conglomerate disappear north of the Missisquoi River in the terrane of the Dunham and Saxe Brook dolomites. They also disappear southward in the internally conformable carbonate-quartzite section in west-central Vermont—at the Lamoille River in western belts and at the Winooski River in eastern belts (Cady, 1945,

p. 537–538). The Parker slate is possibly cut out southward at the unconformity (p. 533) beneath the Rugg Brook formation. Similarly the St. Albans slate is probably cut out to the south (p. 538) at the unconformity that extends from beneath the Mill River conglomerate. Therefore, the original lateral gradation of these slates into the carbonate-quartzite section of west-central Vermont cannot be proved or disproved. The Skeels Corners slate extends southward beneath surficial deposits at the Lamoille River, where it apparently grades laterally southward into the carbonate-quartzite section in the west limb of the Hinesburg synclinorium (S. W. Stone, 1951, Ph.D. Thesis, Harvard Univ., p. 17). In the east limb of this synclinorium, however, the Skeels Corners slate and also the Rockledge conglomerate extend south of the Winooski River where they appear to be interbedded with the carbonate rocks (Cady, 1945, p. 537).[6]

The slate-conglomerate terrane, which in the west limb of the St. Albans synclinorium between the Missisquoi and Lamoille rivers is not more than 20 miles from north to south, is much more extensive in eastern areas (Cady, 1945, p. 538); conglomerate is, however, less abundant to the east. The spread of this terrane into the east limb of the Hinesburg synclinorium is one bit of evidence.[7] A better indication of the eastward expansion, particularly of the slate, is the widespread distribution of its correlatives—the Oak Hill slate and the Sweetsburg slate—in the northern extension of the Enosburg Falls anticline in Quebec.

The mixed carbonate-quartzite and graywacke-shale assemblages of the St. Albans synclinorium and the Enosburg Falls anticline grade northeastward into rocks that are predominantly of the graywacke-shale assemblage but contain interbeds of quartzite, dolomite, and limestone. Cooke and the author jointly traced critical units into the area north of the upper Yamaska River (Cooke, 1954, p. 37, 39), where some become indistinguishable in a terrane chiefly of quartz-sericite-chlorite phyllite and schist. The phyllite and schist overlie the Tibbit Hill volcanics (Cambrian?), which are at the axis of the Enosburg Falls anticline and are in turn topped by the Sweetsburg slate (Middle Cambrian). The Pinnacle formation grades north into quartzite, the West Sutton slate passes locally into phyllite and schist, the Gilman becomes less and less quartzitic and only locally distinguishable, and the Oak Hill slate apparently pinches out. The White Brook and Dunham dolomites and the Scottsmore quartzite are represented by thin tongues or intermittent lenticular masses that are interbedded in the phyllite and schist. P. H. Osberg (1958, written communication) lumps the rocks of the phyllite-schist terrane north of the upper Yamaska River in a single stratigraphic unit referred to by him as the "Bonsecours formation" (Cambrian? and Lower Cambrian).

Graywacke-Shale Assemblage of the Green Mountain Anticlinorium and Vicinity

GENERAL STATEMENT: The rocks of the Green Mountain anticlinorium (Cambrian?, Cambrian, and Ordovician) are principally metamorphic derivatives of the graywacke-shale assemblage. North of the Yamaska

[6] The interpretation of the relationship of the interbedded slates and conglomerates of northwestern Vermont to the carbonate-quartzite section of west-central Vermont presented here disagrees with that of Shaw (1958). Shaw's stratigraphic diagram of western Vermont (Fig. 3, p. 533) shows unconformities, which are not well documented, between Monkton quartzite, "Rugg Brook" (Winooski) dolomite, and Danby formation in the area south of the Lamoille River. His provision that "The apparent conformity of the Winooski and the underlying Monkton quartzite can be explained by reworking of the upper beds of the exposed Monkton . . ." (p. 536) is unsatisfactory. The Monkton grades up into the overlying Winooski by interlamination (Cady, 1945, p. 530, 531, 532, 533). Reworking to produce pseudoconformability would destroy transitions such as that between the Monkton ad Winooski made up of interbedded quartz sandstone and carbonate rock, though it might leave an apparent transition in an entirely quartz sandstone section.

[7] Slate in the eastern part of the Hinesburg synclinorium was considered all of Cambrian age and thrust westward over Lower Ordovician limestones of the east limb of the synclinorium as a thin slice dragged up beneath the Hinesburg overthrust (Cady, 1945, p. 574, Pl. 10). Studies completed in the fall of 1959, however, have shown that most of the "slate" south of the latitude of the city of Burlington is calcareous phyllite that overlaps gradationally westward on the upper strata (probably Bascom Zone 4) of the Lower Ordovician limestones. Lateral gradation from phyllite to limestone is very clear. The phyllite is therefore of Early Ordovician rather than of Cambrian age, and there is no thrust beneath it. Plate 1 of this paper was prepared for publication before the gradational relationship was discovered and shows the thrust slice as it appears in the earlier publication (Cady, 1945, Pl. 10).

River in Quebec, the terrane of this assemblage extends into a wide belt west of the anticlinorium in which the rocks are either unmetamorphosed or are only slightly metamorphic. The unmetamorphosed rocks are distinguished by interbeds of red slate that are rarely recognizable among the metamorphosed rocks of the axial region of the anticlinorium, to the east. Red slates are, moreover, rarely found anywhere in the graywacke-shale terrane in Vermont, where metamorphism was generally more intense than in Quebec. The metamorphic derivatives of the red slate are chiefly light-green quartz-sericite-chlorite phyllite and schist. Metamorphism presumably caused the reduction of the ferric iron (in hematite) in the slates to ferrous iron, most of which entered iron-bearing silicates in the phyllite and schist. The unmetamorphosed rocks west of the Green Mountain anticlinorium in Quebec are the principal clue to the protoliths of the rocks of the metamorphosed terrane.

UNMETAMORPHOSED TERRANE WEST OF THE GREEN MOUNTAIN ANTICLINORIUM IN QUEBEC: The section of the graywacke-shale assemblage that is west of the Green Mountain anticlinorium north of the Yamaska River in Quebec (northern extension of the St. Albans synclinorium) is more than 7000 feet[8] thick (Pl. 2, sec. A-A′). This section makes up the Quebec group (Cambrian, Lower and Middle Ordovician), which is typically developed in the vicinity of the City of Quebec (Logan, 1863, p. 225; Osborne, 1956, p. 169–170). The principal rocks are red shale, green graywacke that includes granules and pebble conglomerates; gray, black, and green siltstone; and gray to black shale and limestone. The red shale and green graywacke (Cambrian and Lower Ordovician), which have been more or less loosely referred to the "Sillery" formation (Melihercsik, 1954, p. 166; Osborne, 1956, p. 169), characterize chiefly the lower two-thirds of the section; they occupy extensive areas in the belt southeast of the St. Lawrence River (Ells, 1888, map; 1896, map) and extend several hundred miles northeast from the vicinity of Granby, Quebec. The gray to black shale and limestone (Middle Ordovician), which have been referred to as the "Farnham limestone" (Ells, 1896, p. 17, 25, 27), are characteristic of the upper third of the section and occupy synclinal tracts immediately northwest of the Green Mountain anticlinorium.

Large masses of limestone (Lower Ordovician?) that crop out in the vicinity of Acton Vale, Quebec (Logan, 1863, p. 243–245, 712–718), are an anomalous feature of the graywacke-shale terrane of the Quebec group. Some are exposed in discontinuous outcrops over areas of at least a square mile. The limestone was referred to as a dolomite by Logan, probably because in many places it contains just enough magnesium carbonate to produce the characteristic buff to gray reticulate weathered surface of dolomite. It is interbedded with various rocks of the Quebec group, including the volcanic rocks, apparently as tonguelike or lenticular zones. These zones range in thickness from a few inches to several hundred feet. At least one laterally persistent thick zone and several other zones of limestone are apparent.

Spectacular boulder conglomerates are well-known features of the Quebec group, particularly in the vicinity of the City of Quebec (Osborne, 1956, p. 181–188). The phenoclasts are commonly angular blocks of limestone and dolomite. The matrix material is most generally dolomite, though calcitic limestone is nearly as common. The boulder conglomerates are abundant in the middle part of the section (Upper Cambrian and Lower Ordovician), above much of the zone that contains green graywacke and red shale. Beds of gray sandstone (Middle? or Upper? Cambrian) are associated with the lowermost of the zones of boulder conglomerate.

Volcanic rocks of mafic to intermediate composition are locally interbedded in the terrane of the Quebec group west of the Green Mountain anticlinorium. Amygdaloidal andesite appears to be the most common. A specimen collected by the author 3 miles northeast of Acton Vale, Quebec, is diabasic to trachitic in texture and contains as essential minerals andesine and augite. Several other specimens that were collected in the vicinity are so thoroughly chloritized and carbonatized that their original composition is indeterminate. Spilitic volcanic rocks are reported about 20 miles southwest of the City of Quebec (Melihercsik, 1954, p. 168). Volcanic rocks were first reported at this and several other local-

[8] This figure is based on thicknesses of formations near the St. Lawrence River, in ascending order: Charny formation, more than 2000 feet; Lauzon formation, more than 2000 feet; Lévis formation, more than 1000 feet; Quebec City formation, approximately 2000 feet (Osborne, 1956, p. 194; Rasetti, 1946, p. 703–704; Raymond, 1913, p. 29).

ities by Logan, who referred to them as "amygdaloidal diorite" (1863, p. 241–244).

The rocks of the Quebec group north of the Yamaska River are probably synchronous with rocks to the south, near the international boundary, that are chiefly of the carbonate-quartzite assemblage and that are included in the combined sections of (1) the Enosburg Falls anticline, (2) the St. Albans synclinorium, and (3) the Philipsburg slice. The principal units in the Quebec group—"Sillery" and Farnham—and also the limestone near Acton Vale are shown in Plate 3 as projected from the north into the three sections at the boundary.

The total thickness of the rocks in the sections at the boundary (8000–10,000 feet) is similar to that of the Quebec group near the City of Quebec (more than 7000 feet). The relationship of the Quebec group to the rocks near the boundary is only partly clear, however, because the Morses Line and Stanbridge slates (=upper Farnham—Middle Ordovician) unconformably overlap the zone in which the lower rocks of the Quebec group ("Sillery" and lower Farnham—Cambrian to Middle Ordovician) join the sections near the boundary. This zone of overlap is chiefly in the Canadian extension of the St. Albans synclinorium. The Morses Line slate on the east limb of the synclinorium lies directly upon the Sweetsburg slate = Skeels Corners and Hungerford slates (Middle Cambrian) of the west limb of the Enosburg Falls anticline (Pl. 2, sec. B-B′).

The principal relationship between the graywacke-shale assemblage of the Quebec group and the synchronous sections to the south that contain the carbonate-quartzite assemblage is probably facies change marked by interlamination of the contrasting types of rock in a zone of lateral transition. Some formations in the sections at the international boundary contain rocks that resemble more or less closely some of the rocks of the Quebec group: volcanic rocks in the Tibbit Hill, graywacke in the Pinnacle formation, purple to red slate in the West Sutton, massive gray quartzite in the Scottsmore and Rugg Brook formations, conglomerates such as those of the Rugg Brook, Mill River, Rockledge, Gorge, and Mystic formations, and the slate of the Morses Line (= Stanbridge) have their counterparts in the Quebec group. Conversely, the limestone that is interbedded in the Quebec group near Acton Vale strongly resembles that of the carbonate-quartzite assemblage near the boundary, especially as represented in the Philipsburg slice, where the resemblance of the St. Armand and Corey formations to the "Acton Vale limestone" is striking and suggests that the latter is a northward-projecting tongue (or tongues) of the rocks exposed in the Philipsburg slice.

The similarities between some of the rocks of the Quebec group and formations in the Enosburg Falls anticline, particularly the West Sutton slate, Gilman quartzite, Oak Hill slate, and Sweetsburg slate, increase northward until it is difficult to differentiate the formations in the Enosburg Falls anticline from the Quebec group. The slates are especially difficult to tell from the Farnham slate. The rocks of the northern extension of the Enosburg Falls anticline are chiefly of the graywacke-shale assemblage and are therefore fundamentally more like the typical Quebec group than are those of the carbonate-quartzite assemblage in the St. Albans synclinorium and the Philipsburg slice. At the St. Francis River, 5.5 miles northwest of Richmond, Quebec, rock referred to by Dresser (Dresser and Denis, 1944, p. 390) as "reddish Sillery slate" and by Ells (1896, p. 24–25) as "purple bands," and that Osberg (1958, oral communication) correlates with the West Sutton slate, crops out in the west limb of the Enosburg Falls anticline. The rarity of red slate in the Enosburg Falls anticline is probably chiefly attributable to increased metamorphic grade in the direction of the axis of the Green Mountain anticlinorium.

GREEN MOUNTAIN ANTICLINORIUM: The graywacke-shale assemblage in the Green Mountain anticlinorium is extremely thick, particularly in Vermont. The section at the international boundary from the lowest rocks (Cambrian?) of the Camels Hump group or Sutton schist that are exposed at the axis of the anticlinorium up to the top of the Moretown or Beauceville formations (Middle Ordovician) on the east limb is estimated to be about 30,000 feet thick. These rocks thicken southward in Vermont to more than 50,000 feet at the Winooski River. (*See* White and Jahns, 1950, p. 190–191.) They probably thin to less than 30,000 feet in Quebec.

The protolith of the metamorphic rocks in the Green Mountain anticlinorium probably included shale (including red shale), siltstone, and graywacke, much like those west of the anticlinorium in Quebec. These have been

metamorphosed to gneiss, granulite,[9] feldspathic quartzite, and slate.

The volcanic rocks are much more abundant in the anticlinorium than in areas to the west and in Vermont are more consistently mafic. The chemical compositions of both the Bolton volcanic rocks in Quebec (Fairbairn, 1933, p. 555–556) and the Belvidere Mountain amphibolite in Vermont are very close to that of Daly's average basalt. The Bolton volcanic rocks, which are interbedded in the Beauceville and Moretown formations, are represented by pillow greenstone and amphibolite (Clark and Fairbairn, 1936, p. 16). The Belvidere Mountain amphibolite, which is at the top of the Camels Hump group, is made up of rather thinly bedded metamorphosed volcanic detritus.

Quartzite is common in some stratigraphic zones, as particularly in parts of the Ottauquechee and Moretown formations in Vermont and in the lower part of the Mansonville formation (=lower "Caldwell") in Quebec. Most distinctive is a gray (carbonaceous) quartzite in the Ottauquechee and lower Mansonville that is intersected by a boxwork of white quartz veins. Much of the quartzite is probably metamorphosed quartz sandstone, but some thin, even, white weathered beds in the Moretown and the Beauceville formations appear to have been originally chert. Feldspathic quartzite grades into granulite. Intraformational conglomerates are common in the Beauceville formation. Conglomerate in the Umbrella Hill formation is basal to the Moretown formation in part of northern Vermont.

The only limestone of any significance is the Melbourne (Middle Cambrian) of Cooke (1954, p. 39–41), a carbonaceous limestone (Dresser, 1906, p. 70) that occurs chiefly in the northern extension of the Cambridge syncline in Quebec. Similar limestone in about the same stratigraphic position occurs in scattered areas in the Cambridge syncline in Vermont (Christman, 1959, p. 38).

Facies transitions are the rule rather than the exception in these rocks. They include changes in both sedimentary and metamorphic features and are therefore complex.

[9] The term granulite refers to "a granulose metamorphic rock, composed of even-sized interlocking granular minerals" (Holmes, 1928, p. 111). The granulites that are referred to here are like those described by Harker (1939, p. 246), which are "rich in quartz" and "contain also a large amount of feldspar."

The Melbourne limestone (Middle Cambrian) in Quebec, which Osberg (1958, written communication) shows as an upper member of the Sweetsburg slate, interfingers with the Sweetsburg (Cooke, 1954, p. 40–41; Dresser, 1906, p. 72–73). A little limestone like that of the Melbourne is interbedded in the Ottauquechee formation in northern Vermont; possibly before erosion from the Green Mountain anticlinorium it was coextensive with the Melbourne limestone.

The quartz-sericite-chlorite schist of the Stowe formation (Upper Cambrian? and Lower Ordovician?) in central Vermont grades northward into granule conglomerate and phyllitic graywacke in the upper part of the Mansonville formation in Quebec. Both the Stowe and the upper Mansonville are commonly greenish, but the upper Mansonville also includes purple to red phyllite and slate that probably reflect lower metamorphic intensity. The purple to red slates and the phyllitic graywacke, which strongly resemble the unmetamorphosed rocks west of the Green Mountain anticlinorium in Quebec, increase in abundance northward along the east limb of the anticlinorium, where they are contained in the "Caldwell facies" (De Romer, 1957, p. 5–6; 1958, p. 5–7).

The Moretown formation (Middle Ordovician) in Vermont grades north into the Beauceville slate and Bolton volcanic rocks in Quebec. The composition of the Moretown ranges locally, both along and across the strike, from that of a rock formed entirely of basaltic rock detritus through highly feldspathic phases (granulite) to pure quartzite, or through chloritic facies to slate, such as that of the Beauceville. The haphazard distribution, especially of volcanic and slate facies, is illustrated west of Montpelier in the area north and south of the Winooski River by the areal pattern of greenstone, amphibolite, and slate (Cady, 1956).[10]

Metamorphic variations are probably responsible for the principal differences between

[10] The unpredictable distribution of sedimentary facies, particularly east of the axis of the Green Mountain anticlinorium, has led to problems of stratigraphic nomenclature. The granulite of the Moretown formation (Cady, 1956; Cady, Albee, and Chidesler, in press; Albee, 1957), as an example, intervenes throughout most of the 40-mile belt between the type locality of the Cram Hill formation (chiefly phyllite and slate) south of the Winooski River (Currier and Jahns, 1941, p. 1493–1496) and synchronous slates mapped as Cram Hill southwest of Lake Memphremagog (Doll, 1951, Pl. 1).

the lower rocks of the graywacke-shale assemblage in the Green Mountain anticlinorium and those west of the anticlinorium in Quebec. The Camels Hump group and Sutton schist (Cambrian? and Cambrian), at the core of the anticlinorium, are synchronous with the lowermost part of the typical Quebec group, west of the anticlinorium, but differ in appearance because red slate and green graywacke, such as occur to the west, are rarely present. The slate and graywacke, respectively, are apparently altered in the anticlinorium to quartz-sericite-chlorite schist and quartz-albite-sericite-chlorite gneiss.

A significant exception to this relationship, which seems nevertheless to demonstrate the premetamorphic lithology of the Sutton schist, is in the St. Sylvestre area, a little southwest of the Chaudière River in the Notre Dame Mountains (Fig. 1). There, red as well as green and gray impure sandstone and siltstone, which are probably part correlatives of the Camels Hump group and Sutton schist, are well within the northwest limb of the Green Mountain anticlinorium (Benoit, 1958, p. 5, map 1214). They are separated from the main body of the red rocks, in the lower part of the typical Quebec group near the St. Lawrence River, by a synclinal tract of gray to black shale (Farnham?) in the upper part of the Quebec group. Red slate and graywacke of the "Caldwell facies" (Upper Cambrian and Lower Ordovician) are very close to the axis of the anticlinorium, where it plunges northeast in the Notre Dame Mountains between St. Magloire and L'Islet (Fig. 1) (Béland, 1957, p. 13, map 1098). This suggests that some of the younger and less deeply buried of the red rocks may, before erosion, have bridged the anticlinorium to the south as well as northeast of St. Magloire.

The upper rocks of the graywacke-shale assemblage, which crop out chiefly in the east limb of the Green Mountain anticlinorium and retain some of their red color, are much like those of the typical Quebec group west of the anticlinorium. The similarity between the Beauceville slate (Middle Ordovician), on the east limb of the anticlinorium, and the Farnham slate (Middle Ordovician), in synclinal tracts west of the anticlinorium, is striking. Both are characterized by gray to black slate that contains brownish-gray silty beds a fraction of an inch thick and 1–4 inches apart. Exposures of the Beauceville approach within 10 miles of the Farnham in the vicinity of Asbestos, Quebec, where the anticlinal axis is depressed (Ells, 1887, map), and the author believes that before erosion they were joined across the axis. The principal facies changes in the Beauceville-Farnham, westward across the axis of the anticlinorium, are depletion of volcanic members and acquisition of gray limestone interbeds.

Carbonate-Quartzite Assemblage of the Lake Memphremagog Synclines and Stoke Mountain Anticline

The rocks in the Lake Memphremagog synclines and on the flanks of the Stoke Mountain anticline (Silurian and Devonian) are chiefly of the carbonate-quartzite assemblage. They total about 1400 feet in the Lake Memphremagog synclines (Cooke, 1950, p. 64–65) and are more than 6000 feet thick (Clark, 1937c, p. 49) northeast of the Stoke Mountain anticline near Lake Aylmer (Fig. 1).

The section in the Lake Memphremagog synclines is chiefly of calcareous slate (Glenbrooke), which is underlain by quartz conglomerate (Peasley Pond) and overlain by dark-gray limestone (Sargent Bay). The Lake Aylmer group contains about the same kinds of rocks, and, as in the sections at Lake Memphremagog, calcareous shale predominates. These rocks are possibly at the eastern border of originally much more extensive rocks of the carbonate-quartzite assemblage that are now eroded from axial regions of the Green Mountain anticlinorium and from belts to the west of the anticlinorium. Lower Devonian limestone preserved in diatreme breccia west of the foreland at Ste. Hélène Island near Montreal (Fig. 1) (Clark, 1944; 1952, p. 81–83) suggests such an original distribution of the carbonate-quartzite assemblage. The nearest stratigraphic sections of post-Ordovician rocks of the carbonate-quartzite assemblage, west and northwest of the Green Mountain anticlinorium, are in the Helderberg Escarpment southwest of Albany, New York (Ruedemann, 1930, p. 41–60), and on Anticosti Island in the Gulf of St. Lawrence (Twenhofel, 1928, p. 54–62). In Plate 3 these sections are projected north and southwest respectively into the foreland section at the international boundary.

Mixed Lithic Assemblages of the Connecticut Valley-Gaspe Synclinorium

Both the carbonate-quartzite and graywacke-shale assemblages are represented in the Connecticut Valley-Gaspe synclinorium, although the graywacke-shale assemblage predominates. The section of these rocks (Silurian and De-

vonian) is 15,000–20,000 feet thick at the international boundary (Doll, 1951, p. 19, 21, 23, 26, 34). A comparable thickness is found in central Vermont (White and Jahns, 1950, p. 191).

The carbonate-quartzite assemblage is represented chiefly by metamorphosed silty limestone and limy siltstone that are interbedded in phyllite and slate of the Barton River formation. The calcareous beds make up less than half of the rock, and for this reason the Barton River is shown in Plate 3 as representative of the graywacke-shale assemblage. Some units, such as the quartzite, conglomerate, and limestones of the Shaw Mountain and Ayers Cliff formations, are almost entirely of the carbonate-quartzite assemblage. The bulk of the graywacke-shale assemblage in the Connecticut Valley-Gaspe synclinorium in Vermont is in the Westmore (=Gile Mountain), which is made up of schist and micaceous quartzite and contains only minor limestone.

Volcanic rocks are relatively rare in the Connecticut Valley-Gaspe synclinorium. Greenstone occurs in the Shaw Mountain formation (Silurian) in the west limb of the synclinorium 2.8 miles S. 25° W. of Albany, Vermont. Pillow amphibolite occurs in the Standing Pond volcanics (Silurian or Devonian or both) of the east limb of the synclinorium in Vermont (Dennis, 1956, p. 22, 75, 76–77; Doll, 1944, p. 17–18). The Standing Pond volcanics pinch in and out along the strike, which is in general northerly; they also pinch out eastward toward the Boundary Mountain anticlinorium.

The Westmore formation and other formations with limestone interbeds form a belted pattern in Vermont, between a wide area to the west that contains limestone and an area to the east that contains very little limestone (Dennis, 1956, Pl. 3; Murthy, 1957, Fig. 2, 3, Pl. 3). It has not been established to what extent the belted pattern is the result of interbedding of rocks of contrasting facies and to what extent it may be attributed to folding of simple stratigraphic units (Eric and Dennis, 1958, p. 58–61; Murthy, 1958). (*See* Dennis, 1959; Murthy, 1959a; 1959b; White, 1959.)

The belt to the east, in Vermont, includes rocks that contain only minor limestone and is wider near the international boundary; northeast of Marbleton, Quebec, the noncalcareous portion of the St. Francis group occupies nearly the full width of the belt of Silurian and Devonian rocks that is centered in the Connecticut Valley-Gaspe synclinorium (Cooke, 1951). The western edge of the noncalcareous terrane enters Quebec from Vermont, about at the transition from the Barton River formation to the Westmore formation (Doll, 1951, Pl. 1). It then extends diagonally northeastward in the direction of Marbleton (Cooke, 1950, p. 30) and includes older rocks, to about as old as the Northfield slate. This relationship lends support to facies interpretations of the belted terranes of calcareous and only slightly calcareous rocks in Vermont.

The lithologic similarity of the Northfield slate, which is near the base of the Silurian and Devonian section in the west limb of the synclinorium, to the Meetinghouse slate, which seems to occupy a similar position on the east limb (west limb of the Boundary Mountain anticlinorium), suggests that the two join at depth in the synclinorium. Possibly the Northfield and Meetinghouse are not a time-stratigraphic unit; they may be a transgressing stratigraphic unit that is younger on the east limb of the Connecticut Valley-Gaspe synclinorium than on the west limb (Pl. 2). If this be the case, the Northfield and Meetinghouse probably grade laterally westward into the Barton River (=? Waits River) formation (Pl. 2). This interpretation explains both the apparently thinner section (approximately 12,000 feet as opposed to 15,000–20,000 feet) on the east limb of the Connecticut Valley-Gaspe synclinorium (Billings, 1956, p. 8, 10–11) and the eastward disappearance of the carbonate rocks in the synclinorium.

Graywacke-Shale Assemblage of the Boundary Mountain Anticlinorium

The rocks of the Boundary Mountain anticlinorium, which total about 15,000 feet thick (Ordovician through Lower Devonian) in northern New Hampshire (Billings, 1937, p. 468), are almost entirely of the graywacke-shale assemblage. Notable exceptions are the Clough quartzite and Fitch (limestone) formation (Silurian) of the east limb of the anticlinorium; in most places these total only 1000 feet or less (Billings, 1956, p. 23, 26).

Rocks of the graywacke-shale assemblage that are beneath the horizons of the Clough and Fitch in northeastern Vermont and northern New Hampshire include the Albee formation and Ammonoosuc volcanics (Middle Ordovician), which are much like the Moretown formation and volcanic rocks interbedded with it on the east limb of the Green Mountain anticlinorium. The section that contains the Albee and Ammonoosuc is a little more than

8000 feet thick in northern New Hampshire (Billings, 1937, p. 468), but inasmuch as its base is not exposed the full thickness is unknown. The Albee consists primarily of slate, phyllite, feldspathic phyllite, quartzose phyllite, argillaceous quartzite, quartzite and feldspathic quartzite, where least metamorphosed (Billings, 1956, p. 16). The Ammonoosuc volcanics range in composition from mafic to felsic, and their protolith includes basalt, andesite, and rhyolite (Billings, 1956, p. 18). The basalt and andesite are metamorphosed to greenstone and amphibolite. The Albee and possibly the Ammonoosuc connect at depth in the Connecticut Valley-Gaspe synclinorium with the Moretown and associated volcanic rocks that crop out in the belt of the Green Mountain anticlinorium.

Rocks of the graywacke-shale assemblage that are above the horizons of the Clough and Fitch formations in the Boundary Mountain anticlinorium and vicinity in northeastern Vermont and northwestern New Hampshire (Silurian? and Devonian) include the Meetinghouse slate and Gile Mountain formation in the west limb of the anticlinorium and the Littleton formation to the east of axis of the anticlinorium. The section that contains the Meetinghouse and Gile Mountain is 9000 feet thick (White and Billings, 1951, p. 651); and that of the Littleton is 5000–15,000 feet thick (Billings, 1937, p. 468; 1956, p. 7), depending upon the depth to which it has been eroded. The thickness of 15,000 feet is comparable to the thickness of 15,000–20,000 feet reported in the section (Silurian and Devonian) in the Connecticut Valley-Gaspe synclinorium, in northern Vermont.

The protolith of the Meetinghouse slate was presumably a black shale. The protoliths of the Gile Mountain and Littleton formations appear to have been chiefly interbedded graywacke and shale that are now metamorphosed to slate, phyllite, schist, and gneiss. The Littleton formation contains interbeds of greenstone and amphibolite, some of which have pillow structure (Billings, 1937, p. 490) and the protoliths of which were probably basalt and andesite (Billings, 1956, p. 28). Before they were eroded the Gile Mountain and Littleton formations probably joined across the axis of the Boundary Mountain anticlinorium.

The Seboomook and Frontenac formations (Devonian), which are a little northwest of the north end of the Boundary Mountain anticlinorium, in the vicinity of Lake Megantic (Fig. 1), are units in Quebec that are apparently equivalent to the Meetinghouse and Gile Mountain and to the Littleton. The Seboomook is composed of "dark gray slates" and "some impure sandstones", and the Frontenac is composed of "greenstones, gray schistose quartzite", and "some gray slates" (Marleau, 1958, p. 2). The greenstones of the Frontenac are amygdaloidal and contain pillow structure.

Stratigraphic Correlation

General Statement

Stratigraphic correlation is most nearly complete in the rocks of the foreland belt, west of the Green Mountain anticlinorium in Vermont. There the sequences of the rocks are better established paleontologically; this is in turn related to the fact that the rocks are little metamorphosed and contain well-preserved fossils. There, also, the carbonate-quartzite assemblage predominates; this assemblage is distinguished by relatively thin but laterally persistent lithologic units or sequences of units. For similar reasons the most successfully correlated rocks east of the foreland belt are of the carbonate-quartzite assemblage.

A special concern of the present paper is to show the correlation of rocks of the graywacke-shale assemblage in the metamorphic terranes east of the foreland. A few distinctive lithologic units in the graywacke-shale terrane indicate such correlation: some persist widely in the belt east of the foreland, some have been traced westward to places where they interfinger in the established sections on the foreland, and one or two contain fossils. The rocks of the graywacke-shale assemblage on the east limb of the Green Mountain anticlinorium correlate with the Cambrian and Ordovician section on the foreland. Those of the Boundary Mountain anticlinorium in turn correlate with the fossiliferous Ordovician to Devonian section on the east limb of the Green Mountain anticlinorium.

Plate 3 shows both north-south correlation across the international boundary and east-west correlation between stratigraphic units in the various north-trending structural belts.

Stratigraphic Equivalence Across the International Boundary

The problem of correlation of rocks north and south of the international boundary is chiefly one of terminology; differences for each structural belt are indicated in Plate 3. Clarification of the relationship of the stratigraphic

terms used in Quebec to those used south of the boundary is necessary because east-west correlation, especially from limb to limb of the Green Mountain anticlinorium in Vermont, requires the tracing of critical units across the depressed axis of the anticlinorium in Quebec.

The equivalence on opposite sides of the international boundary of stratigraphic units in sections of the Champlain-Richelieu foreland, the Highgate Springs-St. Dominique slice, and the Philipsburg slice has been shown in the literature.[11] The stratigraphic equivalents north and south of the boundary in the St. Albans synclinorium, Enosburg Falls anticline, and Connecticut Valley-Gaspe synclinorium are also, with a few minor exceptions, clearly shown in the literature[12]. The author's field studies have further clarified the correlation of the Scottsmore quartzite and Sweetsburg slate, respectively, of the Enosburg Falls anticline in Quebec with the Rugg Brook formation and the Skeels Corners and Hungerford slates of St. Albans synclinorium and Enosburg Falls anticline, south of the international boundary.

The stratigraphy of the rocks of the east limb of the Green Mountain anticlinorium, in an area crossed by the Winooski River about 50 miles south of the international boundary, has recently been described by the author (Cady, 1956). The stratigraphic units in the Winooski River section have been traced northward in the anticlinorium and in the Connecticut Valley-Gaspe synclinorium, in the course of field work by the author and his associates, Albee (1957; 1957, Ph.D. thesis, Harvard Univ.) and A. H. Chidester. (*See* Cady, Albee, and Chidester, in press.)

The Camels Hump group, which is composed chiefly of schist, quartzite, and gneiss, is at the core of the Green Mountain anticlinorium and lowest in the section in north-central Vermont. Formations that overlie the Camels Hump group on the east limb of the anticlinorium are: the Ottauquechee, which is composed of interbedded carbonaceous quartzite and phyllite; the Stowe, which is made up principally of interbedded quartz-sericite-chlorite schist and greenstone; and the Moretown, whose most characteristic rock is a finely laminated granulite-quartzite. The Moretown formation underlies the Shaw Mountain formation, which is included in Plate 3 with the stratigraphic units of the Connecticut Valley-Gaspe synclinorium.

The stratigraphic section on the east limb of the Green Mountain anticlinorium is somewhat modified near the international boundary. The Belvidere Mountain amphibolite (Keith and Bain, 1932, p. 173–174) appears at the top of the Camels Hump group (Albee, 1957; 1957, Ph.D. thesis, Harvard Univ., p. 28–30). The Ottauquechee formation and the lower part of the Stowe formation acquire granule conglomerate and beds of coarse graywacke. The Umbrella Hill formation, which contains a cobble conglomerate, separates the Stowe from the Moretown (Albee, 1957; 1957, Ph.D. thesis, Harvard Univ., p. 45–48). The Moretown contains a thick interstratified section of mafic to intermediate volcanic rocks that in turn contain interbeds of siltstone.

The author has traced the Camels Hump group, the Ottauquechee, Stowe, and the Moretown formations, respectively, northward into the Sutton (Clark, 1934, p. 6, 9–11; Fortier, 1946, Ph.D. thesis, Stanford Univ.), the Mansonville (Clark, 1934, p. 11; Fortier, 1946, Ph.D. thesis, Stanford Univ.), and the Beauceville (Cooke, 1950, p. 33) formations in Quebec. The Ottauquechee formation becomes the lower part of the Mansonville of Clark, which De Romer (1957, p. 4–5; 1958, p. 4–5), in the area west of Mt. Oxford, refers to as "rocks transitional between the Sutton and Caldwell facies." The Stowe formation, which becomes the upper Mansonville, is designated as the "Caldwell facies" by De Romer (1958, p. 5–7). The volcanic rocks in the Moretown formation are continuous with the Bolton volcanic rocks, which Ambrose (1957, p. 166–169) has found are interbedded in the Beauceville formation in Quebec.

The stratigraphy of the rocks of the Cambridge syncline, which is in the west limb of the Green Mountain anticlinorium, is being studied, and correlation between Vermont and Quebec along this syncline should soon become clear. Carbonaceous schist, quartzite, and

[11] *See* Clark (1952, p. 42–46), Clark and Strachan (1955), Erwin (1957), Hawley (1957), Kay (1937, p. 275–276), and Oxley (1950) for the sections in the Champlain-Richelieu foreland; Kay (1945; 1950; 1958), Kay and Cady (1947), and Logan (1863, p. 273–280) for the sections in the Highgate Springs-St. Dominique slice; and McGerrigle (1930, p. 184–186) for the section in the Philipsburg slice.

[12] *See* Clark (1934, p. 6–8) for the section in the St. Albans synclinorium; Booth (1950, p. 1135–1136, 1138) for the section in the Enosburg Falls anticline; and Dennis (1956, p. 23–32), Doll (1943a, p. 59; 1951, p. 17, 22, 23), Ells (1887, p. 9), Hitchcock (1877, p. 17–18), Logan and Hall (1865), and Logan and Hunt (1855, p. 52–53) for the sections in the Connecticut Valley-Gaspe synclinorium.

limestone, which were discovered several years ago in the town of Cambridge, Vermont, by the author and his associates, resemble the Sweetsburg slate and Scottsmore quartzite, in nearby parts of Quebec (Pl. 3), and the Melbourne limestone of Cooke (1954, p. 39–41), which is in the upper part of the Sweetsburg (Osberg, 1958, written communication) in the northern extension of the Cambridge syncline in Quebec.

In the Boundary Mountain anticlinorium, the stratigraphic equivalence of formations in Vermont (and northern New Hampshire) and Quebec has been suggested especially in the publications of geologists who have done field work in Quebec (Ells, 1887, p. 31; McGerrigle, 1935, p. 76–77; Marleau, 1958, D.Sc. thesis, Laval Univ., p. 31, 63–65, 95–97). Current studies by J. C. Green and N. L. Hatch in Northern New Hampshire and by A. L. Albee in adjacent Maine are clarifying the stratigraphic relationships, and the results of their field work (1959, oral communications) support correlation of the Albee and Ammonoosuc in New Hampshire with the Arnold River formation of Marleau (1958, D.Sc. thesis, Laval Univ., p. 32) in Quebec. The correlation shown in Plate 3 (annotation 21) is tentative pending completion of the work of these men.

East-West Stratigraphic Correlation Between Structural Belts

GENERAL STATEMENT: East-west stratigraphic correlation depends chiefly upon the lateral persistence of lithologic units that are repeated in adjacent thrust slices or on the opposite limbs of folds, or that project without change between strata whose facies vary laterally and lack distinctive features. This dependence upon laterally persistent units is most critical in and east of the Green Mountain anticlinorium, where broad anticlinal and synclinal tracts must be bridged, facies variations are general, stratigraphic units monotonously thick, and fossils have nearly all been destroyed by metamorphism.

BELTS WEST OF THE GREEN MOUNTAIN ANTICLINORIUM: A distinctive formation of the structural belts west of the Green Mountain anticlinorium is the Beldens, which is composed of interbedded limestone and dolomite. The Beldens formation is reported from deep wells in the Champlain-Richelieu foreland (Belyea, 1952, p. 11–14; Clark, 1955, p. 10), and it crops out in the Highgate Springs-St. Dominique slice (Kay, 1945; 1950; 1958, p. 78–80; Kay and Cady, 1947). It is of Early Ordovician age (Kay, 1950; 1958, p. 79), and the author has correlated it lithologically with the St. Armand and Corey formations of the Philipsburg slice. The Gorge formation of the St. Albans synclinorium is useful in correlation between thrust slices; the dolomite strata in this formation are lithologically identical with the Rock River dolomite of the Philipsburg slice and the Clarendon Springs dolomite in southern areas of the foreland (Cady, 1945, p. 537). The author has found that the Morses Line slate at the top of the section in the St. Albans synclinorium can be traced into the lithologically similar Stanbridge slate at the top of the section in the Philipsburg slice.

Stratigraphic correlation between sparsely fossiliferous rocks of the east limb of the St. Albans synclinorium (west limb of the Enosburg Falls anticline) and the better established section in the west limb of the synclinorium depends chiefly upon recognition of distinctive lithologic features of the Dunham dolomite, the Rugg Brook formation, and the Mill River and Rockledge conglomerates. The interbedded slate formations—Parker, St. Albans, Skeels Corners, and Hungerford—all look much alike and are indistinguishable except where their sequential relationship to the Dunham, Rugg Brook, Mill River, and Rockledge is clear. The Dunham is mainly buff-weathering dolomite that is locally characterized by undulant thin siliceous bedding laminae; in some places it is sandy. The Rugg Brook is made up chiefly of dolomite, which is difficult to distinguish from that of the Dunham but contains beds of dolomitic conglomerate and "almost pure, gray weathering sandstone" (Howell, 1938, p. 100). The gray sandstone, which is commonly crisscrossed by veins of white quartz, is a distinctive feature of the Scottsmore quartzite of the Enosburg Falls anticline in Quebec (Clark, 1936b, p. 148–149). The Mill River conglomerate and the Rockledge conglomerate, which contain limestone phenoclasts in a sandy limestone matrix, are indistinguishable from each other (Shaw, 1958, p. 542), and either or both possibly occur in the east limb of the St. Albans synclinorium (Pl. 3, annotation 5). The Rockledge is more extensive in the established section in the west limb of the synclinorium (Shaw, 1958, Pl. 1), which suggests that it rather than the Mill River occurs in the east limb.

Lithologic similarities suggest that the Scottsmore quartzite of the west limb of the Enosburg Falls anticline in Quebec may form a bridge between the Middle Cambrian Rugg Brook formation of the St. Albans synclinorium in Vermont and some sandstones and quartzites of the Quebec group, north of the Yamaska River in Quebec. The Scottsmore resembles gray sandstones, some of which have a siliceous cement, that are interbedded in the Quebec group. Osborne (1956, p. 181) reports gray sandstone near the City of Quebec (Fig. 1) associated with limestone conglomerate that is about halfway between the base and the top of the typical Quebec group. The sandstone and conglomerate are in the Lauzon formation of Rasetti (1946, p. 700–702). The Scottsmore also resembles the Kamouraska quartzite (Dresser, 1912, p. 14–20), which is interbedded in the Quebec group along the south shore of the St. Lawrence River (Ells, 1888, p. 69) northeast from the vicinity of L'Islet (Fig. 1). The Kamouraska contains interbeds of limestone conglomerate. The conglomerates of the Lauzon and Kamouraska differ from those of the Rugg Brook formation chiefly in that most of the boulders or blocks are of limestone rather than dolomite as in the Rugg Brook.

Paleontologic evidence seems to deny a close correlation of the Rugg Brook and the Scottsmore with the Lauzon and the Kamouraska. The Lauzon contains fossils that are Late Cambrian and younger (Rasetti, 1946, p. 700); the youngest fossils in the boulders of the Kamouraska have been considered of Late Cambrian age (p. 702). The lower 220 feet of the Lauzon have not yielded fossils and are possibly of Middle Cambrian age. The *Hungaia magnifica* fauna in the boulders of the Kamouraska is possibly of early Late Cambrian age (A. B. Shaw, 1958, written communication).

GREEN MOUNTAIN ANTICLINORIUM: The Camels Hump group in Vermont and the equivalent Sutton schist in Quebec, which are at the base of the section in the axial anticlines of the Green Mountain anticlinorium, probably correlate in time and in part lithologically with the Tibbit Hill to Oak Hill and Parker formations, inclusive, in the section in the Enosburg Falls anticline (Pl. 3, annotations 2, 3). This interpretation, which approximates that of Clark (1934, p. 9–10), has become clearer as the stratigraphic relationships of the overlying Ottauquechee formation and Mansonville slate have become better understood.

The section in the Cambridge syncline in northern Vermont, which intervenes between the axial anticlines of the Green Mountain anticlinorium and the Enosburg Falls anticline, contains chiefly the undifferentiated rocks of the "Jay Peak series" of Booth (1950, Pl. 1). The "Jay Peak series" probably correlates principally with the upper rocks of the section in the Enosburg Falls anticline, as well as with the upper parts of the Camels Hump group and Sutton schist, in the axial anticlines of the Green Mountain anticlinorium. Correlative rocks in Quebec are included chiefly in the "Bonsecours" formation of Osberg (1958, written communication).

The Ottauquechee formation (=lower Mansonville) contains interbeds of massive carbonaceous quartzite, chiefly near its base, that are similar to quartzite in the Rugg Brook formation in the Enosburg Falls anticline and the St. Albans synclinorium in northwestern Vermont. The quartzite in the Ottauquechee is also like quartzite, which is possibly the Scottsmore, in the Cambridge syncline in Vermont. The carbonaceous phyllite of the Ottauquechee is comparable to the St. Albans, Skeels Corners, and Hungerford slates (=Sweetsburg) west of the axial anticline of the Green Mountain anticlinorium. These similarities suggest a correlation which has been substantiated by Osberg (1956), who has traced the Sweetsburg slate through a depression in the Green Mountain anticlinorium a little west of Asbestos, Quebec. The Sweetsburg and underlying Scottsmore quartzite become the lower Mansonville and Ottauquechee formations of the sections in the east limb of the axial anticline of the anticlinorium near the international boundary.

The Sweetsburg continues northeastward in the Green Mountain anticlinorium, beyond the depression near Asbestos (Cooke, 1954, p. 43). There Cooke and the author were also able to trace the Scottsmore quartzite. Interbedded gray quartzite and black phyllite or schist, such as those of the Scottsmore and Sweetsburg, respectively, are common features of the northeastern extension of the anticlinorial belt in the Notre Dame Mountains (Fig. 1). Some similar rocks, which are near the axis of the anticlinorium and are possibly beneath correlatives of the Tibbit Hill volcanics, the Pinnacle formation, and the White Brook dolomite (Benoit, 1958, p. 4, Map 1214; 1958, D. Sc. thesis, Laval Univ., p. 27–31, 33–38, Map 1), apparently correlate with parts of the Sutton schist and Camels Hump group near the international boundary. The uppermost of the interbedded massive gray quartzites and black phyllites, however, which are principally on the flanks of the anticlinorium, have been assigned not only to the Scottsmore and Sweetsburg or Ottauquechee (=lower Mansonville) but to the L'Islet formation of Dresser (1912, p. 20-22; "lower Sillery" of Ells, 1888, p. 63-64) and to the Rosaire group of Béland (1952, p. 4; 1954a, p. 3; 1954b, p. 3; 1957,

p. 13-15). (See also Fortier, 1946, Ph.D. thesis, Stanford Univ.)

The L'Islet formation, as mapped in a reconnaissance study by Dresser (1912, Map 34A), passes around the northeast-plunging nose of the Green Mountain anticlinorium in the Notre Dame Mountains northeast of St. Magloire, Quebec (Fig. 1). There, according to Béland (1957, p. 15), the L'Islet "more or less includes both the Rosaire and the Caldwell groups as . . . distinguished" by detailed mapping. Dresser's lithologic description of the L'Islet, however (1912, p. 20-21), shows that he referred chiefly to the rocks now included in the Rosaire.

Dresser (1913, p. 51) mentioned that "the Georgia (lower Cambrian) emerges from beneath the L'Islet near the boundary line of the state of Vermont." He apparently referred to the section that is topped by the Scottsmore quartzite and the Sweetsburg slate in the west limb of the Enosburg Falls anticline in Quebec. Ells (1896, p. 57) mentions "hard sandstones with gray and black slates . . . resembling the lower Sillery" (=L'Islet) at Bethel, 11 miles southwest of Richmond, Quebec, on the west limb of the Green Mountain anticlinorium. Dresser (Dresser and Denis, 1944, p. 389) assigned to the L'Islet, near St. Felix de Kingsey 7 miles northwest of Richmond (also on the west limb of the anticlinorium), rocks that Osberg (1958, written communication) maps as Sweetsburg. Knox (1917, p. 234) reported the L'Islet formation on the east limb of the anticlinorium southwest of Thetford Mines (Fig. 1). Farther south on the east limb, near the international boundary, Fairbairn (1932, Ph.D. thesis, Harvard Univ., p. 40) correlated the Mansonville, presumably the lower Mansonville (=Ottauquechee), with the L'Islet.

The Stowe formation (=upper Mansonville) possibly correlates with part of the quartz-sericite-chlorite schist in the Cambridge syncline, west of the axial anticline of the Green Mountain anticlinorium, though it seems more likely that the Stowe, if once present, has been eroded from this belt. Rocks that resemble those of the Ottauquechee formation and that are tentatively correlated with the Ottauquechee seem to occupy an axial position in the Cambridge syncline (Christman, 1959, p. 46-47); if they do they are probably at the top of the section that remains, and the Stowe is gone. This is certainly true in the northern extension of the Cambridge syncline in Quebec where Osberg (1958, written communication) maps the Sweetsburg (=Ottauquechee) at the axis of the syncline.

Rocks that are the time equivalent of the Stowe formation are probably not present on the west limb of the Enosburg Falls anticline in Vermont and nearby parts of Quebec. There the Middle Ordovician Farnham formation, including the Morses Line slates (=upper Farnham), lies on the Skeels Corners-Hungerford and Sweetsburg slates (=Ottauquechee), and rocks of Late Cambrian and of Early Ordovician age are missing (Pl. 2). This overlap of the Farnham formation on the Sweetsburg slate continues northward on the west limb of the Enosburg Falls anticline nearly to the latitude of Richmond, Quebec. Eight miles west of Richmond, massive graywacke and granule conglomerate like that of the "Caldwell facies" (=upper Mansonville) is mapped by Osberg (1958, written communication) stratigraphically above the Sweetsburg on the west limb of the Green Mountain anticlinorium. The "Caldwell facies" cannot be traced eastward through the depression in the anticlinorium near Asbestos, as can the Sweetsburg, because it is eroded. However, its lithology and its sequential relationship to the Sweetsburg (=Ottauquechee and lower Mansonville) confirm the correlation with the "Caldwell facies" on the east limb of the anticlinorium.

The typical "Caldwell" (MacKay, 1921, p. 20–24) is on the east limb of the Green Mountain anticlinorium, northeast of the depression near Asbestos. It is synonymous with the "Caldwell facies" of De Romer (1957, p. 5–6; 1958, p. 5–7) near the international boundary on the east limb of the anticlinorium. The "Caldwell" contains green, red, and gray graywacke and slate that are like the rocks of various formations that have been recognized on the west limb of the anticlinorium, including the Sillery of Dresser (1912, p. 22–25; "upper Sillery" of Ells, 1888, p. 64) and the Armagh group of Béland (1952, p. 4; 1954a, p. 3; 1954b, p. 2–3; 1957, p. 11–13). (See also Fortier, 1946, Ph.D. thesis, Stanford Univ.) Béland (1957, p. 13) indicates that the relationship of the "Caldwell" and the Armagh on opposite limbs of the anticlinorium in the St. Magloire area (Fig. 1) "suggests that the Armagh is . . . contemporaneous with the Caldwell," though some possible evidence to the contrary had been previously turned up on an area northeast of the St. Magloire (Béland, 1954b, p. 4). Work by Benoit (1958, D.Sc. thesis, Laval Univ., p. 106) in the area east of St. Sylvestre seems to confirm the correlation of the Armagh with the "Caldwell."

"Sillery" is reported in several places on the west limb of the anticlinorium: Osberg (1956) refers to "Sillery" east of Acton Vale, 8 miles west of Richmond, stratigraphically above the Sweetsburg slate (Osberg, 1958, written communication). Dresser (Dresser and Denis, 1944, p. 389, 390) designates as "Sillery" rocks that are stratigraphically beneath the Sweetsburg slate, about 6 miles northeast of this locality in the town of Kingsey (5.5 miles northwest of Richmond). This "Sillery" correlates with the West Sutton slate of the section in the Enosburg

Falls anticline near the international boundary (Osberg, 1958, written communication). Ells (1896, p. 24, 58) reports "Sillery", also in the town of Kingsey but 5 miles north of the Dresser locality, that is stratigraphically above the Sweetsburg as mapped by Osberg. Benoit (1958, D.Sc. thesis, Laval, Univ., p. 38–47) designates a unit called "Oak Hill-Armagh" in the vicinity of St. Sylvestre (Fig. 1) that contains rocks of Armagh type. With these rocks are interbedded rocks that are partly like, and presumably correlate with, those of the West Sutton, Gilman, and Sweetsburg of the Enosburg Falls anticline near the international boundary. Ells (1888, p. 62) recognized "Sillery" in the "Oak Hill-Armagh" terrane about 3 miles north and also 10 miles southwest of St. Sylvestre. Ells (1888, p.86) also described the Sillery at the type locality of the Armagh, 15 miles northwest of St. Magloire (Béland, 1957, map 1097).

"Sillery" has also been reported in at least two places on the east limb of the Green Mountain anticlinorium: Knox (1917, p. 234–235) reported "Sillery" stratigraphically above the L'Islet southwest of Thetford Mines (Fig. 1). Dresser (Dresser and Denis, 1944, p. 391) remarks that "remnants of the formation are found at least as far south as Shipton Pinnacle . . . and it may continue as far as Mount Orford" (Pl. 1).

The "Sillery" and "Armagh" that are beneath the horizon of the Sweetsburg slate on the west limb of the Green Mountain anticlinorium are probably synchronous with the Lower Cambrian Charny formation (Rasetti, 1946, p. 698–700) in the typical Quebec group, near the City of Quebec. Benoit (1958, D.Sc. thesis, Laval Univ., p. 107-109) has suggested that the rocks of the St. Sylvestre area that correlate with the section in the Enosburg Falls anticline (Oak Hill series of Clark, 1936b) and contain rocks of the Sillery and Armagh type correlate with the Charny. Higher rocks in the St. Sylvestre area that contain red slates like those of the Sillery and also limestone conglomerate are tentatively correlated with the Lauzon and Levis formations in the middle and upper part of the typical Quebec group (Benoit, 1958, D.Sc. thesis, Laval Univ., p. 107). Correlation of the Sillery and Armagh on the west flank of the Green Mountain anticlinorium in Quebec with the Charny is difficult to prove because they are separated from the typical Charny by synclinal tracts of younger rocks, chiefly correlatives of the Farnham slate, in the upper part of the Quebec group west of the anticlinorium.

The Beauceville formation (=Moretown) probably correlates with the Farnham formation at the top of the Quebec group west of the Green Mountain anticlinorium in Quebec (Pl. 3). The two were probably a single continuous formation before dissection of the anticlinorium (Pl. 2). The Beauceville was included in the Farnham by Dresser (1912, p. 27–28; 1913, p. 25; Dresser and Denis, 1944, p. 404). Osberg (1956) correlated the Beauceville with the Stanbridge, which is equivalent to at least the upper part of the Farnham. "Greyish and bluish-grey-striped massive slates" (Ells, 1888, p. 62–63), like those of the Beauceville south of Shipton Pinnacle, crop out in the upper part of the Quebec group (probably Farnham) in the St. Sylvestre area (Fig. 1) to the northwest of the anticlinorium and suggest a correlation.

BELTS EAST OF THE GREEN MOUNTAIN ANTICLINORIUM: One of the most distinctive formations east of the axis of the Green Mountain anticlinorium in Vermont is the Moretown, which is characterized by a finely laminated granulite-quartzite that is commonly referred to as "the pinstripe." A similar "pinstripe" also appears in the Albee formation of the Gardner Mountain anticline (axial anticline of the Boundary Mountain anticlinorium) and suggests a correlation (Pl. 3). Correlation of the Albee with the Moretown is further suggested by the fact that the Albee is succeeded conformably by the Ammonoosuc volcanics, which are similar to mafic to intermediate and more rarely felsic volcanic rocks that are interbedded in the upper part of the Moretown and in the Beauceville slate in Quebec.

Correlation of the Albee and Ammonoosuc with the Moretown was originally proposed by W. S. White (1946, Ph.D. thesis, Harvard Univ., p. 129, 131–132). Later, White and Billings (1951, p. 649–662) placed the Albee and Ammonoosuc above the Waits River formation, thereby precluding correlation of the Albee and Ammonoosuc with the Moretown. This stratigraphic reassignment of the Albee and Ammonoosuc was in part based upon the discovery in the Waits River formation of poorly preserved fossil cup corals that were tentatively assigned an Ordovician age pending discoveries of better preserved fossils or of clearer stratigraphic and structural relationships (Cady, 1950a). Stratigraphic and structural relationships that support a Silurian and possibly Devonian age for the Waits River formation and its correlatives in Quebec (Pl. 3, annotation 22) (Boucot and others, 1953) are now becoming increasingly evident. Billings (1956, p. 98) has recently suggested, as an alternative interpretation, that the Albee and Ammonoosuc may be older than the Waits River and that they correlate with the Moretown and Cram Hill formations, respectively. This interpretation is substantiated by correlation of rocks that overlie the Albee and Ammonoosuc with those that succeed the Moretown and Beauceville.

Correlation of rocks that overlie the Moretown with those that overlie the Beauceville seems clear. The Moretown formation is directly overlain by the Shaw Mountain formation on the west limb of the Connecticut Valley-Gaspe synclinorium in Vermont (Cady, 1956; Currier and Jahns, 1941, p. 1496–1501; Doll, 1951, p. 18–20). Where the Shaw Mountain formation is absent the Northfield slate lies directly, and to all appearances conformably, on the Moretown formation. The Beauceville formation is overlain by the Peasley Pond conglomerate in the Lake Memphremagog synclines (Clark, 1936a; Cooke, 1950, p. 64–65). The Glenbrooke shale succeeds the Peasley Pond much as the Northfield follows the Shaw Mountain.

Lithic features and sequential relationships indicate that the Shaw Mountain and Northfield correlate with the Peasley Pond and Glenbrooke. The Shaw Mountain formation is a tan-weathered sericite-quartz-chlorite-albite phyllite-schist that grades near the base into schistose quartz conglomerate and near the top into phyllitic, blue-gray, locally crinoidal limestone with rust spots. It is comparable to the Peasley Pond, which is a schistose quartz conglomerate that contains interbeds of phyllite-schist and is topped by rusty weathered limestone beds. The Northfield slate contains thin calcareous interbeds and is like the Glenbrooke shale, though more metamorphosed.

Similar rocks are found in the Lake Aylmer group, near the northeast end of the Stoke Mountain anticline. Three-quarters of a mile northwest of East Angus, Quebec, on the east limb of the Stoke Mountain anticline, the author discovered rocks like those of the Shaw Mountain and Peasley Pond formations. These rocks are on strike with the Shaw Mountain formation on the west limb of the Connecticut Valley-Gaspe synclinorium in Vermont, but their connection with the Shaw Mountain formation in the southern area is covered by intervening surficial deposits. Near East Angus, these rocks are included both in areas that Cooke (1951) mapped as "conglomerate, with interbedded grit and quartzite" in the Lake Aylmer group, and in other areas that he apparently unintentionally placed in the St. Francis group. The conglomerate in the area northwest of East Angus contains pebbles of rhyolite from the adjoining Beauceville group in the Stoke Mountain anticline, which Cooke included in the "Sherbrooke group" (Pl. 3, annotation 16).

"Argillaceous" types of limestone (Cooke, 1950, p. 75) in the Lake Aylmer group southeast of Marbleton, Quebec, at the northern tip of the Stoke Mountains, resemble both the Glenbrooke shale and the Northfield slate. Moreover, they grade southeastward into the St. Francis group of the Connecticut Valley-Gaspe synclinorium. This gradational relationship is like that of the Glenbrooke shale to the overlying Sargent Bay limestone and that of the Northfield slate to the Waits River formation (=Ayers Cliff and Barton River).

The Clough quartzite, which is of Early-Silurian age and overlies the Albee and Ammonoosuc on the east limb of the Gardner Mountain anticline (axial anticline of the Boundary Mountain anticlinorium) in New Hampshire, is nearly synchronous with the Lower(?) or Middle Silurian Peasley Pond conglomerate of the Lake Memphremagog synclines. Currier and Jahns (1941, p. 1497) remarked on the lithic similarity between the conglomerate of the Clough quartzite and the basal conglomerate of the Shaw Mountain formation, although a correlation was not intended. Later, White (1946, Ph.D. thesis, Harvard Univ., p. 131–132) proposed a correlation of the Clough with the Shaw Mountain formation. More recently, Boucot and others (1953) have suggested the equivalence of the Clough and the Shaw Mountain on the basis of field studies in southern Vermont and New Hampshire and in nearby Massachusetts.

Billings (1956, p. 94–97) has summed up most of the arguments for and against correlation of the Clough with the Shaw Mountain and the Albee and Ammonoosuc with the Moretown and Cram Hill. He shows that the outcome of the controversy hinges to a considerable extent on whether or not the Shaw Mountain formation correlates with the Peasley Pond conglomerate and whether or not the succeeding Northfield slate and Waits River formation, respectively, correlate with the Glenbrooke shale and Sargent Bay limestone. The author's field evidence, already described, is that the Shaw Mountain and Peasley Pond correlate, and that a correlation of the Clough with the Shaw Mountain is therefore tenable.

A major hindrance to correlation of the Clough with the Shaw Mountain is the failure of quartz conglomerate and quartzite to appear on the west limb of the Boundary Mountain anticlinorium. This limb of the Boundary Mountain anticlinorium is between the type locality of the Clough on the east limb of the Boundary Mountain anticlinorium and that of

the Shaw Mountain on the west limb of the Connecticut Valley-Gaspe synclinorium. The stratigraphic zone in question is at the "Monroe Line" (Pl. 1), which is between the Meetinghouse slate to the west and various formations, namely, the Orfordville, Albee, and Ammonoosuc to the east, in the Boundary Mountain anticlinorium. Absence of the quartz conglomerate and quartzites, such as those of the Shaw Mountain and Clough, can be variously explained by the Monroe thrust (Billings, 1956, p. 97; White, 1946, Ph.D. thesis, Harvard Univ., p. 131), by unconformable overlap (Billings, 1956, p. 97–98), by failure to be deposited, by erosion, or by combinations of these explanations. Facies variations in the west limb of the Boundary Mountain anticlinorium are a possible explanation. This is a pertinent explanation suggested by scarcity or lack of quartz conglomerate in the Shaw Mountain formation (Cady, 1956). However, the general absence in the west limb of the Boundary Mountain anticlinorium of rocks, such as the quartzite, that are comparable to those of either the Shaw Mountain or Clough suggests that rocks of the time-stratigraphic interval of the Shaw Mountain and Clough are missing and either were never deposited or were eroded before deposition of the succeeding Meetinghouse slate, which overlaps them.

Billings (1956, p. 97–98) has suggested that if, as the author believes, the tops of the beds are to the west along the "Monroe Line," unconformable overlap by the Meetinghouse slate might be an alternative to the "Monroe fault." Such an overlap onto the Albee and Ammonoosuc seems a very likely explanation of the failure of rocks correlated with the Shaw Mountain and Clough to appear on the west limb of the Boundary Mountain anticlinorium (Pl. 2). Overlap extinction not only of correlatives of the Shaw Mountain and Clough but also of the Albee, where it is missing east of the "Monroe Line," implies a deeply penetrating unconformity. Such a profound unconformity would be unnecessary if it could be shown that the Orfordville and its associated volcanic rocks are facies of the Albee and Ammonoosuc.

The evidence for the "Monroe fault" is, according to the original interpretation, principally structural (Eric, White, and Hadley, 1941). Three quarters of a mile north of the village of Monroe, New Hampshire, "bedding and flow cleavage in both the Meetinghouse and the Albee" are reported to "strike at small angles toward a breccia occupying the contact plane" (Eric and Dennis, 1958, p. 55). Regionally, however, the principal evidence is a stratigraphic hiatus in which parts of the section east of the "Monroe Line" are apparently missing. Thus, at the type locality of the "Monroe fault" just discussed, the Ammonoosuc volcanics are absent and the Albee formation adjoins the contact, whereas south of the vicinity of Woodsville, New Hampshire, the Orfordville formation adjoins the contact, and both the Albee and Ammonoosuc are absent (White and Billings, 1951, p. 670). In view of these regional relationships, and because the only reported exposure of the contact is that north of Monroe, the evidence for a fault in other than the type locality is primarily stratigraphic. Therefore, demonstration of the thrust depends upon the vicissitudes of stratigraphic interpretation.

Lyons (1955, p. 133) has found little structural evidence for this fault west of the Connecticut River near White River Junction, Vermont; instead, the contact is presumably sedimentary. He points out that the cleavage-bedding and drag-fold readings and, more significantly, lava pillows, near the critical contact, indicate that rocks to the west are stratigraphically above those to the east. The author found these cleavage-bedding and drag-fold relationships identical with some he had previously recorded in the outlet tunnel of the flood-control dam at Union Village, Vermont, farther north along the "Monroe thrust." The evidence from both of these localities is therefore consistent with a position on the west limb of an anticline. Lyons (1955, p. 134) concludes, however, chiefly on stratigraphic grounds, that the tops of beds are to the east: presumed younger beds (Orfordville) are to the east of the critical sedimentary contact ("Monroe Line") in the vicinity of White River Junction.

The problem of the correlation of the Meetinghouse slate and Gile Mountain formation on the west limb of the Boundary Mountain anticlinorium with the Littleton formation on the east limb is closely tied to the interpretation of the direction of the tops of beds along the "Monroe Line." If the tops of beds are to the west, the Meetinghouse and the Gile Mountain are, as the author believes, on the west limb of the anticlinorium and, like the Littleton, stratigraphically above the Albee and Ammonoosuc. If the tops are to the east, the whole section to the east of the "Monroe Line," which includes the Littleton, is stratigraphically above the Meetinghouse and Gile Mountain.

The equivalency of the Gile Mountain with the Littleton was first indicated by Doll (1944, p. 28). Later, W. S. White (1946, Ph.D. thesis, Harvard Univ., p. 131–132) suggested correlation of the Meetinghouse and Gile Mountain with the Littleton as a possible alternative to placing them in a stratigraphic position above the Littleton. Billings (1956, p. 98) points out

that interpretation of the Meetinghouse and Gile Mountain, as well as of the Standing Pond volcanics and the Waits River formation, as part of a western facies of the Littleton, is especially attractive in southeastern Vermont, where the Littleton and Gile Mountain appear to be in contact without any structural break.

The principal objections to this interpretation seem to arise from failure to appreciate that lateral variations, particularly across the strike, are the rule rather than the exception in geosynclinal deposits such as those represented by the Meetinghouse, Gile Mountain, and Littleton. One objection cites the failure of the Standing Pond, which is probably a fairly reliable time-stratigraphic unit, to persist laterally east or west of the vicinity of the eastern part of the belt of the Waits River formation. The disparate thicknesses and grain sizes of clastic rock assemblages that are probably facies equivalents in eastern and western belts have also been cited as objections. Such objections seem to the author to have little validity, except in the discussion of foreland sections, like that of the Champlain Valley, where lithology and thickness are fairly dependable tools of correlation.

Geologic Age

General Statement

The stratified rocks that are to the west of the axis of the Green Mountain anticlinorium are of Cambrian(?), Cambrian, and Ordovician age (Booth, 1950, p. 1136–1137; Cady, 1945, p. 524–525; Clark, 1934, p. 5–9; 1955, p. 7; Shaw, 1958, p. 522–523). Those to the east range in age from Cambrian to Devonian (Billings, 1948, p. 50; Clark, 1923; 1934, p. 9–13; Logan, 1863, p. 427–437). Ages based on fossils have been fairly easily assigned to rocks of the carbonate-quartzite assemblage and to representatives of the graywacke-shale assemblage that are interbedded with or are in the vicinity of rocks of the carbonate-quartzite assemblage. This is probably because the carbonate-quartzite assemblage bespeaks a more stable tectonic setting and, therefore, a more favorable life environment and one in which the fossils are less likely to be either diluted by inorganic clastic constituents or obliterated by regional metamorphism.

Fossils contained in the carbonate-quartzite assemblage and associated rocks represent one or more faunal zones in each of the series of the Cambrian and Ordovician, probably in each of those of the Silurian, and in the Lower and Middle Devonian. Terranes entirely of the graywacke-shale assemblage that are in and east of the Green Mountain anticlinorium have yielded fossils diagnostic of only Middle Ordovician and Early Devonian age.

Cambrian(?), Cambrian, and Ordovician

Rocks at the base of the section in the Green Mountain anticlinorium in northern Vermont and adjacent Quebec are assigned the age of Cambrian(?). They lie conformably beneath fossiliferous lower Cambrian rocks but contain no fossils.[13] The formations of Cambrian(?) age that are in conformable section with the fossiliferous Lower Cambrian are the Tibbit Hill, Call Mill, Pinnacle, White Brook, and West Sutton of the Enosburg Falls anticline (Pl. 3). The lower part of the "Jay Peak series" in the Cambridge syncline and the Camels Hump group and Sutton schist of the axial anticlines of the Green Mountain anticlinorium, all of which probably correlate with the rocks in the lower part of the section in the Enosburg Falls anticline, are also assigned a Cambrian(?) age.

Two recently reported determinations of Precambrian age that were based on lead-isotope ratios of ores should be mentioned before closing the discussion on Cambrian(?) age. Wilson, Russell, and Farquhar (1956, p. 553, 554) report Precambrian ages of 1 billion ± 200 million years, which they have extrapolated over wide areas of the Appalachians including the Green Mountains and Sutton Mountains. The ores used were from prospects in the vicinity of St. Magloire, Quebec, and from the Suffield mine southeast of Sherbrooke, Quebec. At St. Magloire the ore, according to Béland (1952, p. 10; 1957, p. 34–42), is in veins in sills that intrude the Middle Ordovician Beauceville formation. At the Suffield mine (Cooke, 1950, p. 132), mineralization is in a schistose rhyolite flow, mapped as part of the "Sherbrooke group" (Cooke, 1951; *see* Pl. 3, annotation 16), that is interbedded in the Beauceville. These age determinations are thus contradicted by paleontologic and stratigraphic evidence in Vermont and Quebec.

[13] This age assignment, which conforms with the practice of the U. S. Geological Survey, differs from that of Clark (1936b, p. 146), who included in the Lower Cambrian all rocks that, regardless of whether they are fossiliferous, lie conformably beneath formations that contain fossils of Early Cambrian age.

The base of the rocks assigned with certainty to the Cambrian is placed beneath the Gilman quartzite (Pl. 3, Enosburg Falls anticline), which is the lowest fossiliferous formation in the stratigraphic section in northern Vermont and adjacent parts of southern Quebec. The single fossil reported from this formation is a brachiopod of the genus *Kutorgina*, which is a Lower Cambrian guide fossil (Clark, 1936b, p. 146). Rocks of each of the epochs of the Cambrian and Ordovician are represented in the belts west of the Green Mountain anticlinorium.[14] The geologic ages of these rocks are amply discussed in existing literature. This literature has grown apace because fossils are fairly common and stratigraphic relationships are easily decipherable.

Assignment of ages to the pre-Middle Ordovician rocks of the Green Mountain anticlinorium depends entirely upon stratigraphic correlations with rocks of known age west of the anticlinorium. Thus, the Scottsmore and Sweetsburg and the Ottauquechee formations (=lower Mansonville) are assigned a Middle Cambrian age because they correlate stratigraphically with the Middle Cambrian section—Rugg Brook to Hungerford, inclusive—in the St. Albans synclinorium (Pl. 3). The author infers from these relationships that the upper parts of the Canels Hump group and Sutton schist are of Early Cambrian age. Also, the upper part of the Mansonville formation (=Stowe) is assigned Late Cambrian(?) and Early(?) Ordovician ages, because it is between the Middle Cambrian lower Mansonville and the overlying Beauceville, which is of Middle Ordovician age.

A corollary of the assignment of a Middle Cambrian age to the Sweetsburg formation is the assignment of a similar age to the Melbourne limestone of Cooke (1954, p. 39–41), which, with some of the Sweetsburg, had been given a Middle Ordovician age (Ells, 1887, p. 17–19; Dresser and Denis, 1944, p. 402, 403). This apparent contradiction seems to arise partly from the fact that the Melbourne and associated rocks, which crop out chiefly near Richmond, Quebec, extend north to the vicinity of Danville, Quebec, where they are within about 2 miles of the main belt of the Middle Ordovician Farnham formation west of the Green Mountain anticlinorium (Ells, 1887, map; 1896, map). Ells (1887, p. 18–19) recognized that the Melbourne limestone underlies rocks that are overlain by the Farnham but attributed the apparent lack of correlation of the Melbourne and Farnham to faulting. Confusion has probably also been produced by the lithic similarity between the interbedded dark limestones and slates of the Farnham and those of the Sweetsburg (including the Melbourne).

A few poorly preserved fossils that are said to suggest a Lower Ordovician age have been reported in the vicinity of the Melbourne limestone 2.5 miles northeast of Richmond (Dresser, 1906, p. 71–72, 74; Dresser and Denis, 1944, p. 402). P. H. Osberg, H. B. Whittington, and the author jointly examined the locality and discovered twiglike secondary features in the schistose limestone that could have been mistaken for the genera of bryozoans (*Stenopora* and *Ptilodictya*) reported. Osberg (1958, written communication) mapped phyllite of the Sweetsburg formation at this locality.

The Beauceville formation, in the Lake Memphremagog synclines in Quebec, has yielded late Middle Ordovician (Trenton) fossils (Cooke, 1950, p. 45–48; Fortier, 1946, Ph.D. thesis, Stanford Univ.). Northeast of Lake St. Francis (Fig. 1), Cooke (1933, p. 44–45) reported Middle Ordovician fossils, older (Normanskill=Chazy) than those in the Beauceville of the Lake Memphremagog synclines.[15] The author has assigned a Middle Ordovician age to the Moretown formation of the east limb of the Green Mountain anticlinorium in Vermont and to the Albee and Ammonoosuc formations east of the Connecticut Valley-Gaspe synclinorium, all of which correlate stratigraphically with the Beauceville.

Rocks of Late Ordovician age are not known in the vicinity of Vermont and neighboring parts of Quebec east of the Green Mountain anticlinorium and are probably not present. The nearest fossiliferous rocks, in southeastern areas, that are known to be of late Ordovician age are in the southeastern limb of the Connecticut Valley-Gaspe synclinorium, about 300

[14] On Plate 3 the Mill River conglomerate, Skeels Corners slate, Rockledge conglomerate, Hungerford slate, and Saxe Brook dolomite are assigned to the Middle Cambrian (Shaw, 1958, p. 548). These formations, all or in part, had been previously considered of Late Cambrian age. Shaw's assignment of a Middle Cambrian age reflects a new correlation of the *Cedaria* faunal zone with elsewhere established sections of the Cambrian rather than a new local interpretation. A. R. Palmer (1959, oral communication), who has the most intimate knowledge of the critical faunas, states that the Middle Cambrian age of the *Cedaria* fauna has not yet been settled; hence the author's assignment of the formations in question to the Middle Cambrian, to conform with the local usage of Shaw, is only tentative.

[15] The rock included in the Beauceville at this locality was later placed by Clark (1937b, p. 42–43) in the St. Francis group, although without explanation. He confirmed a Middle Ordovician age.

miles northeast of the Vermont-Quebec boundary near Matapedia, Quebec (Dresser and Denis, 1944, p. 296–297).

Silurian and Devonian

Fossils diagnostic of each epoch from Early Silurian through Middle Devonian are reported on and east of the east limb of the Green Mountain anticlinorium.[16] They occur chiefly in rocks of the carbonate-quartzite assemblage, although many are known in the graywacke-shale assemblage. The fossiliferous formations include the Glenbrooke shale and Sargent Bay limestone of the Memphremagog synclines, the Lake Aylmer group of the Stoke Mountain anticline, the St. Juste group of the Connecticut Valley-Gaspe synclinorium, the Seboomook formation at the north end of the Boundary Mountain anticlinorium, and the Clough quartzite, Fitch, and Littleton formations of the east limb of the Boundary Mountain anticlinorium.

The Shaw Mountain formation, in the west limb of the Connecticut Valley-Gaspe synclinorium, and its correlatives, the Peasley Pond conglomerate of the Lake Memphremagog synclines and conglomerates in the basal part of the Lake Aylmer group, near the north end of the Stoke Mountain anticline, are of Early or Middle Silurian age. If these rocks are synchronous with the Clough quartzite, as seems likely, they are of Early Silurian age.

Middle Devonian fossils are reported only from the St. Juste group in Quebec, northeast of the area covered by Plate 1 (Clark, 1923; Gorman, 1955, p. 5). South of the international boundary, the youngest fossiliferous rocks, which are in the Littleton formation, are of Early Devonian age. The Littleton formation is intruded by the New Hampshire plutonic series, which according to radiometric age determinations (296 ± 29 million years, Larsen method) is probably of Middle Devonian age (Lyons and others, 1957, p. 532–534). Therefore, the Littleton formation probably does not contain strata younger than Early Devonian.

[16] Fossils of Early Silurian age are reported by Boucot and Thompson (1958); Middle Silurian by Bancroft (1915, p. 66), Billings (1956, p. 90–92), and Harvie (1914, p. 214); Late Silurian by Boucot (1957, oral communication); Early Devonian by Billings (1956, p. 92–94), Boucot (1953 Ph.D. thesis, Harvard Univ.), Clark (1937c, p. 50–51; 1942), and Laverdière (1936, p. 37–38); and Middle Devonian by Clark (1923).

Some of the most troublesome problems of geologic age are concerned with fossils in rocks chiefly of the graywacke-shale assemblage and probably of Silurian and Devonian age. Most controversial have been the fossils from the Waits River and Gile Mountain formations and to a lesser extent those from the Shaw Mountain formation and the Northfield slate. These formations have been assigned ages that range from Ordovician to Devonian. Faunal investigations have furnished inconclusive and partly conflicting results because of the poor preservation of the fossils. Bucher (1953a, p. 282–283; 1953b) and Boucot (1953) have discussed the preservation of some of these fossils and their usefulness in age determinations.

Currier and Jahns (1941, p. 1500–1501) collected fragments of crinoid stems believed to be of "Middle Ordovician or later age" (p. 1510) from the Shaw Mountain formation. Richardson reported graptolites from the Northfield slate and Waits River formation (Richardson, 1916, p. 142–145; 1919, p. 49–51; Richardson and Camp, 1919, p. 114–115), specimens that were believed by Ruedemann (1947, p. 63) to be of both Deepkill and Normanskill (Early and Middle Ordovician) age. Cady (1950a) collected cup corals from the Waits River formation that were tentatively assigned a Middle Ordovician age but with the provision (p. 495) that "... future discoveries of fossils, or of new stratigraphic and structural relations, could result in changes in age assignments." Clark reported graptolites (1934, p. 12) with "a Trenton aspect" (1949, written communication) from the "Tomifobia slates and limestones" about a mile north of the international boundary; these rocks project across the boundary into the Barton River formation of Vermont. Doll (1943a) collected cystoid and crinoid calyces that show evidence of Silurian or possibly Devonian age from the Westmore formation (Doll, 1951, p. 33–37), which is probably identical with the Gile Mountain formation. Doll (1943b) also reports a brachiopod from the Gile Mountain formation, probably a *Spirifer* of Early Devonian age.

The author and others (Currier and Jahns, 1941, p. 1505–1506) have examined numerous specimens of the "graptolites" from the localities referred to by Richardson, also specimens collected by Richardson, and find that they are streaks of sericite chiefly in the crystalline limestone of the Waits River formation (=Barton River formation). D. J. McLaren and L. M. Cumming (1953, written communications), paleontologists of the Geological Survey of Canada, report as follows on the graptolites collected by T. H. Clark from the extension of the Barton River formation in Quebec and submitted for identification and age determinations by A. J. Boucot:

"Locality: Two miles east of Stanstead [Quebec] from the Tomifobia formation. Locality numbers 9D1, 1OE1, 10F1, 12E1.

"A re-examination of the material considered by Clark to contain graptolites of Trenton age allows no positive conclusions to be drawn. Markings suggestive of highly sheared graptolites occur on the rock samples from all localities but it is impossible to be sure that they are in fact graptolites. A specimen from locality 12E1 contains a dendroid marking resembling *Dictyonema* sp. In my opinion there is no evidence to suggest an age for these specimens" (D. J. McLaren). "Specimens . . . show, on bedding surfaces, anastomosing bands of lighter colored surficial material. There is no evidence of positive graptolite structure, but the markings might be interpreted as sheared distal portions of very elongate graptolite colonies. Definite age determinations cannot be made from this material" (L. M. Cumming).

Boucot (1953, written communication) adds that these reports ". . . make it clear that the age of the Tomifobia formation is still a moot point . . . . Clark's material is probably graptolitic in nature, which leads us to believe that this part of the Barton River is probably no younger than Late Silurian."

The equivocal faunal evidence for the age of the post-Shaw Mountain rocks of the Connecticut Valley-Gaspe synclinorium, particularly the Northfield and Waits River formations, cited above, is tipped in favor of Silurian and Devonian ages, as opposed to an Ordovician age, by the weight of stratigraphic evidence already pointed out for the Early and possibly Middle Silurian age of the Shaw Mountain formation. A Silurian and Devonian age for the rocks above the Shaw Mountain is further substantiated by reassessment of Logan's evidence for like age and correlation of the St. Francis group; this evidence (Pl. 3, annotation 22) has been revived by recent workers, notably Gorman (1954, p. 3–4; 1955, p. 4–5).

Geotectonics

General Statement

Geotectonics is concerned primarily with regional structural geology and secondarily with stratigraphy. As used here it applies to various large tectonic units, which the author has referred to as geotectonic elements (Cady, 1950b). The largest of these approach subcontinental dimensions and contain several systems of stratified rocks. They are related in space and time, respectively, to critical types and sequences of both diastrophic and magmatic events. Interpretation of the diastrophic events is based chiefly upon knowledge of the distribution and sources of deposits of clastic sediments and upon the kinds of stratigraphic breaks between them.

The geotectonic setting of the region of northern Vermont and nearby parts of southern Quebec is the once mobile belt of the Appalachian orthogeosyncline, which trends northeast at the eastern margin of North America. This belt, which includes the anticlinoria and synclinoria as well as the adjacent parts of the foreland to the west, borders the southeastern margin of the stable continental block or craton[17]. The craton includes the regions of the Canadian Shield and the Andirondack Mountains in which the Precambrian basement is exposed and connected stable shelves. The orthogeosyncline and craton are together the major features of geotectonic significance. At the time of its formation the orthogeosyncline was a belt chiefly of subsidence, as evidenced by its thick content of bedded rocks that represent the Cambrian(?) and most of epochs of the Cambrian, Ordovician, Silurian, and Devonian.

The belt of the Appalachian orthogeosyncline, in the region of northern Vermont and southern Quebec, includes several subsidiary tectonic elements that form the immediate setting of the stratified rocks. Most of the rocks are embraced in two parallel and adjoining longitudinal zones of subsidence: the eugeosynclinal zone, which was most mobile, lies to the east, and the less mobile miogeosynclinal zone is to the west, next to the craton (Pl. 2). Various types of second- and third-generation geosynclines are superimposed, not only upon the orthogeosyncline, but upon adjoining parts of the craton.

Although subsidence was the outstanding diastrophic event in the early history of the orthogeosyncline, it was gradually superseded by uplift and finally by folding. Local uplift, chiefly within the eugeosynclinal zone, accompanied subsidence and provided the subaerial sources of most of the clastic sediments, principally those of the graywacke-shale assemblage. General stabilization of the western part of the orthogeosyncline, at the end of the Ordovician, resulted in eastward migration of the miogeosynclinal zone—east of the present Green Mountains and Sutton Mountains—and marked the beginning of consolidation of the orthogeosyncline with the craton. Localized uplift, east of the Green Mountains and Sutton Mountains at this time, is marked by uncon-

[17] Foreland as used here refers only to the mobile shelf, which is marginal to the craton.

formities that are referred to the Taconic disturbance. Folding and uplift of the contents of the orthogeosyncline, after the Early Devonian, is shown by angular unconformities that are referred to the Acadian and Appalachian orogenies. These unconformities mark the completed consolidation of the orthogeosyncline with the craton.

Critical magmatic features (Stille, 1940, p. 13–23; 1950) that are closely related to the geotectonic development of the region began with, accompanied, and followed the diastrophism. They have assisted in the interpretation of geotectonic relationships of the bedded rocks. (*See* Cady and Chidester, 1957.)

Geotectonic Elements

GENERAL STATEMENT: The geotectonic elements are tectonic units that are controlled by essentially constant diastrophic influences. Such units are geosynclines in or near the orthogeosynclinal belt or stable shelves upon the craton; their duration commonly encompassed several periods of geologic history. Diastrophic influences are reflected by the sedimentary content of the geotectonic elements. In the Vermont-Quebec region, for example, the carbonate-quartzite assemblage, which characterizes the miogeosynclinal zone, and the graywacke-shale assemblage, characteristic of the eugeosynclinal zone, both include rocks of Cambrian to Devonian age.

Unconformities are a clue to the distribution of geotectonic elements inasmuch as they mark abrupt changes in diastrophic conditions from subsidence to uplift and possibly folding. Thus, the base of the section in the miogeosynclinal zone in the east limb of the Green Mountain anticlinorium, where Lower Silurian rocks transgress on the eugeosynclinal zone, is marked by an unconformity. West of the mountains, however, at the margin of the foreland, where the Cambrian and Ordovician rocks of the miogeosynclinal zone interfinger with those of the eugeosynclinal zone to the east and lateral gradation is more apparent, the contrasting facies of contained rocks are the principal clue to the boundary between the miogeosynclinal zone and the eugeosynclinal zone (Pl. 2). Because of the commonly gradational relationship between rocks of the eugeosynclinal and miogeosynclinal zones, which preclude separation by geographic barriers, these elements are here referred to as zones within the orthogeosyncline. (*See* Stille, 1940, p. 653–654, 656.)

The eugeosynclinal and miogeosynclinal zones differ genetically in that the former was characterized by prolonged episodes of subsidence and the latter by fairly continuous episodes of stillstand. The most satisfactory evidence of subsidence in the eugeosynclinal zone, in Vermont, is the great thickness of contained rocks compared with the thickness of the section in the miogeosynclinal zone. The thickness of the eugeosynclinal section (Cambrian?, Cambrian, and Ordovician) east of the Green Mountains (to the top of the Cram Hill = upper Moretown) is four to eight times that of the synchronous section in the miogeosynclinal zone west of the Green Mountains (Cady, 1945, p. 524–525; White and Jahns, 1950, p. 100–191). The great prevalence of limestone, dolomite, and orthoquartzite in the miogeosynclinal zone is evidence of stillstand there.

The geotectonic elements recognized in the New England-Maritime-Quebec region, in addition to the eugeosynclinal and miogeosynclinal zones, are exogeosynclines, epieugeosynclines, and taphrogeosynclines. The latter contain rocks that are separated from underlying rocks, principally of the eugeosynclinal zone, chiefly by angular unconformities.

EUGEOSYNCLINAL ZONE: The eugeosynclinal zone is by definition the first folded and more mobile zone of the orthogeosynclinal belt, and it is characterized by volcanic rocks that are chiefly mafic (Stille, 1940, p. 653–654). It includes most of the anticlinoria and synclinoria in the region of northern Vermont and southern Quebec and contains the extremely thick section of rock of the graywacke-shale assemblage.

The distribution of the eugeosynclinal zone in this region is about the same as that of its contained volcanic rocks, most of which include mafic pillow lavas and are probably marine deposits. The graywacke-shale assemblage is, to a lesser extent, a reliable guide to the distribution of the eugeosynclinal zone. The principal feature of the distribution of this zone is the general offlap to the east away from the foreland of its Silurian and Devonian rocks, with respect to the miogeosynclinal zone (Marleau, 1958, D. Sc. thesis, Laval Univ., p. 109).

Another less general and more anomalous feature is an encroachment of the eugeosynclinal zone seemingly upon the foreland north of the Yamaska River in Quebec. This distribution is indicated by the occurrence in the Quebec group of this area not only of widespread rocks of the graywacke-shale assemblage but of interbedded volcanic rocks (Pl.

2, sec. A-A′). The graywacke-shale terrane reaches the northwestern margin of the orthogeosynclinal belt along the St. Lawrence River at and northeast of the City of Quebec.

This relationship is exceptional in that a miogeosynclinal zone that contains rocks synchronous with those of the graywacke-shale terrane is missing between the craton and the eugeosynclinal zone. Possibly the rocks of the eugeosynclinal zone are overthrust from the southeast, and the miogeosynclinal zone is buried beneath the thrust slices. The thrust hypothesis does not, however, explain how coarse sediments in the graywacke-shale terrane, which were derived from the craton (Osborne, 1956, p. 197), traversed such a miogeosynclinal zone.

Northwestward widening of the eugeosynclinal zone and, conversely, the narrowing of the miogeosynclinal zone, in Quebec, have generally been explained since the days of Logan (1861, p. 381; 1863, p. 233) by overthrusting of the eugeosynclinal rocks from the southeast (Osborne, 1956, p. 170, 196). A thrust interpretation seems to be supported by the relationships in and near the Taconic Range (Fig. 1) in western New England and eastern New York (Cady, 1945, p. 567–569). There rocks of eugeosynclinal aspect, some of which are lithologically similar to the graywacke-shale terrane in Quebec, and all of which are of about the same Cambrian to Ordovician time span, lie on Middle Ordovician rocks of the miogeosynclinal zone. Zen (1956) indicates that the relations require either thrusting or overturning upon the Middle Ordovician rocks[18].

The author (1945, Fig. 5, p. 563, Fig. 6, p. 568, 569) and Kay (1937, Pl. 5, p. 286) suggested that the graywacke-shale assemblage ("Sillery") northwest of the Green Mountain anticlinorium in Quebec and the rocks of the Taconic Range and vicinity were coextensive before erosion and were parts of a single and very extensive allochthone that had been thrust from the southeast. This suggestion now seems unlikely because the "Sillery" is in stratigraphic sequence with the Sweetsburg slate beneath and the Stanbridge (= Farnham) slate above (Osberg, 1956), both of which slates may be traced southward into synchronous and autochthonous carbonate, quartzite, and shale strata of western Vermont.

In a modified version of this theory, the allochthone is postulated to be an "early Middle Ordovician thrust plate that was buried by later Middle and Upper Ordovician beds" (Henderson, 1958). The thrust fault cited is that described by Cooke (1954, p. 41–43) beneath the "Caldwell." The author has been unable to find such a fault, and existence of the postulated thrust plate is extremely doubtful.

MIOGEOSYNCLINAL ZONE: The miogeosynclinal zone is a relatively narrow mobile shelf that is near the cratonal margin of the orthogeosynclinal belt and is characterized by the lack of volcanic rocks (Stille, 1940, p. 15, 656) and the predominance of rocks of the carbonate-quartzite assemblage. It is best developed on the foreland in Vermont and immediately adjoining parts of southern Quebec, where it contains Cambrian(?), Cambrian, and Lower and Middle Ordovician rocks (Pl. 1).

Here it contains both continuous sections of the carbonate-quartzite assemblage and mixed sections in which the carbonate-quartzite and graywacke-shale assemblages interfinger. The mixed assemblages increase toward the east and northeast as the transition to the eugeosynclinal zone is approached (Pl. 2). This transition is in the west limb of the Enosburg Falls anticline in Vermont and nearby Quebec, but north of the Yamaska River in Quebec, where the eugeosynclinal zone encroaches on the foreland, the zone of transition passes westward across the St. Albans synclinorium. Tongues or intermittent lenticular masses of the carbonate-quartzite assemblage that extend north of the Yamaska River into the eugeosynclinal zone probably reflect brief episodes of somewhat greater stability in the midst of otherwise mobile and characteristically eugeosynclinal conditions. Because of the evidence for a widespread and unbroken eugeosynclinal zone, persistent shelves or stable nuclei probably could not account for these tonguelike or lenticular occurrences of the carbonate-quartzite assemblage.

The miogeosynclinal zone overlaps the eugeosynclinal zone east of the axis of the Green Mountain anticlinorium. Silurian and Lower Devonian rocks of mixed carbonate-quartzite and graywacke-shale assemblages that are exposed chiefly in the west limb of the Connecticut Valley-Gaspe synclinorium are contained in the overlapping portion of the miogeosynclinal zone. The Silurian and Lower Devonian rocks of the east limb of this synclinorium are nearly all of the graywacke-shale assemblage and are contained in the eugeosynclinal zone southeast of the overlap.

The original westward extent of the miogeosynclinal zone of the Silurian and Devonian is

[18] Zen (1959, p. 4–5) has discussed the folding and overthrust relationships of the Taconic rocks more fully and in the light of more recent field studies.

uncertain chiefly because the contained rocks have been eroded from the axial region of the Green Mountain anticlinorium and from the foreland belt to the west of the anticlinorium. The nearest rocks of Silurian or Devonian age are on stable shelves on the margin of the craton west of the anticlinorium near Albany, New York, and to the northwest on Anticosti Island in the Gulf of St. Lawrence.

EXOGEOSYNCLINE: The exogeosyncline is a secondary geosyncline that is superimposed on rocks of the miogeosynclinal zone and also of the adjacent craton (Pl. 2) (Kay, 1951, p. 17–20, 107). It contains Upper Ordovician rocks of the graywacke-shale assemblage, which are free of volcanic rocks. The eastern extent of the exogeosyncline is indeterminate, inasmuch as the contained rocks are eroded from the eastern part of the foreland. The exogeosyncline probably did not extend east of the axis of the later formed Green Mountain anticlinorium, inasmuch as Upper Ordovician rocks are unknown in Vermont and neighboring parts of Quebec, east of the anticlinorium. The exogeosyncline is thus a discrete trough, rather than a zone within a larger geosyncline, that is filled with rocks only of the graywacke-shale assemblage, and that is centered approximately on the foreland belt.

EPIEUGEOSYNCLINES: The epieugeosynclines are secondary geosynclines that are superimposed on rocks of the eugeosynclinal zone (Kay, 1951, p. 107). They contain Mississippian, Pennsylvanian, and Permian rocks that are chiefly of the graywacke-shale assemblage. Included in the epieugeosynclines are terrestrial volcanic rocks. Epieugeosynclines are known only in regions southeast of northern Vermont and the neighboring parts of Quebec (Kay 1951, p. 56–57), probably because their contained rocks have been eroded from northwestern areas. The nearest rocks that are in an epieugeosyncline are the Moat volcanics of late Permian(?) age (Lyons and others, 1957, p. 540) in the White Mountains of New Hampshire (Fig. 1) (Billings, 1956, p. 35–37).

TAPHROGEOSYNCLINES: The taphrogeosynclines are fault troughs that are filled with sedimentary rocks (Kay, 1951, p. 107). The contents of the troughs include principally rocks of the graywacke-shale assemblage; they are of Late Triassic age. Mafic volcanic rocks are commonly interbedded. Taphrogeosynclines are known only in regions south and southeast of the region of the present paper (Kay, 1951, p. 60). The nearest rocks contained in a taphrogeosyncline are in the Newark group in the Connecticut Valley near Greenfield, Massachusetts (Fig. 1) (Emerson, 1917, p. 91–105, Pl. 10). The Newark group probably extended over at least southern Vermont and New Hampshire before it was eroded to its present distribution. Postmetamorphic faults such as the Ammonoosuc thrust in northern New Hampshire and the Adirondack border faults in the Champlain Valley are comparable with the high-angle faults that border the fault troughs in the Connecticut Valley and suggest that taphrogeosynclines once existed in the region of northern Vermont and southernmost Quebec.

Sources of Clastic Sediments

GENERAL STATEMENT: The principal sources of the clastic sediments deposited in the orthogeosynclinal belt and in nearby areas were within this belt. Tectonic islands and, to some extent, archipelagic volcanic accumulations, which appeared at various times in the eugeosynclinal zone, supplied most of the detritus for rocks of the graywacke-shale assemblage. Boulders of limestone and dolomite were provided by intermittent emergence and erosion of the miogeosynclinal zone. The craton was initially a source of "dirty," coarse clastic sediments and contributed to the graywacke-shale assemblage, but with reduction of its relief the craton supplied only "clean" quartz sands which are characteristic of the carbonate-quartzite assemblage in the adjacent miogeosynclinal zone.

GRAYWACKE-SHALE: The sediments of the graywacke-shale assemblage were derived chiefly from within the eugeosynclinal zone. In Vermont and nearby Quebec most of this assemblage is cut off from a cratonal source by synchronous rocks of the carbonate-quartzite assemblage that occupy much of the miogeosynclinal zone (Pl. 2). Islands in the eugeosynclinal zone were doubtless the source of much of the sediment in the graywacke-shale assemblage. Rocks of the graywacke-shale assemblage in the eastern part of the miogeosynclinal zone that are continuous with those of the eugeosynclinal zone were probably derived from uplifts in the latter zone. The contents of the exogeosyncline were eroded from a very extensive land area uplifted to the east, in the eugeosynclinal zone, at the time of the Taconic disturbance.

Some thick sections of the graywacke-shale assemblage within the eugeosynclinal zone are

difficult to explain *e.g.*, the section in Vermont, which is 10–20 miles thick (White and Jahns, 1950, p. 190–191) and contains a tremendous volume of sediments. Deeply penetrating unconformities, which one would expect to mark the sites from which these sediments were eroded, are rare within the pre-Silurian portion of this section.

An unknown part of the sediments in the eugeosynclinal zone was contributed by volcanic accumulations that extended above the surface of the sea and were intermittently replenished from below. Such additions, which were accompanied by subsidence of the eugeosynclinal zone, were probably the principal source of sediments in some of the thicker sections.

Volcanic constituents are much reduced in the Middle Silurian and later graywacke-shale assemblages of the eugeosynclinal zone; therefore, tectonic islands were probably the principal source of their sediments. Many tectonic islands are probably required to explain the wide distribution of these clastic rocks, which, before erosion, probably extended southeastward to the Atlantic Ocean.

Some of the sediments of the graywacke-shale assemblage that are in the eugeosynclinal zone came from the direction of the craton. These are chiefly sediments that are in rocks near the base of the stratigraphic section and at the western margin of the eugeosynclinal zone. The western source of the graywacke sediments of the Cambrian(?) Pinnacle formation seems clear from the westward increase of grain size as well as volume of typical facies (Booth, 1950, p. 1156–1157). Lower Cambrian rocks of the graywacke-shale terrane northwest of the Green Mountain anticlinorium in Quebec also contain sediments that were derived from the northwest. Osborne (1956, p. 178) says of the cratonal source of some of the sediments of the typical Quebec group:

> "The minerals of the sandstone [Charny] are unquestionably ultimately derived from a terrane similar to that of the Laurentian uplands north of [the City of] Quebec: feldspars can be recognized that are characteristic of the gneisses and of intrusives inclusing anorthosite. On the other hand there is little material derived from rocks of the Grenville series."

Detrital grains of microcline perthite are fairly widespread not only in the typical Quebec group, but also in the Pinnacle and Mansonville formations in southernmost Quebec and in the Ottauquechee and Stowe formations in northern Vermont. These indicate a source terrain in which were exposed felsic plutons older than any such plutons known in the eugeosynclinal zone and probably of Precambrian age. Boulders of microcline granite, sodic granite, syenite, and quartz syenite that are in conglomeratic interbeds in the Pinnacle formation (Booth, 1950, p. 1143) reflect the lithology of these plutons. The eastern limits of distribution of the clastic sediments that were derived from the west and northwest is not clear chiefly because the Cambrian(?) and Cambrian rocks in which they occur are at depth east of the east limb of the Green Mountain anticlinorium. A western source of the sediments in the graywacke-shale assemblages of the upper part of the Camels Hump group and of the Ottauquechee and Stowe formations in Vermont is likely only in the northernmost part of the State. Farther south in Vermont, carbonate rocks and clean quartzite that are above the West Sutton formation and in the miogeosynclinal zone intervene between the eugeosynclinal zone and the margin of the craton to the west.

Ubiquitous distribution of the graywacke-shale assemblage in the orthogeosynclinal belt, characteristic especially of the Lower Cambrian in Quebec, seems at first anomalous. Possibly, however, this is a feature of early stages of all orthogeosynclines before extremes of both subaerial and submarine relief in the vicinity of the margins of the craton were reduced by denudation and deposition. The relief in the vicinity of the City of Quebec appears to have persisted a little longer than that near and south of the international boundary.

CONGLOMERATE: In the western part of the orthogeosynclinal belt, limestone, dolomite, and quartzite boulders occur in conglomerates and also as isolated boulders in finer-grained rock. The boulders were derived from rocks of the carbonate-quartzite assemblage contained chiefly in the miogeosynclinal zone. Some of these source rocks are also interbedded with the graywacke-shale assemblage in the neighboring parts of the eugeosynclinal zone.

The conglomerates and isolated boulders in the western part of the orthogeosynclinal belt are those of the following formations, chiefly in the miogeosynclinal zone: Middle Cambrian Rugg Brook, Mill River, and Rockledge; Upper Cambrian Gorge; Lower Ordovician Highgate; and Middle Ordovician Mystic (=Corliss) and Hathaway. Those chiefly in the eugeosynclinal zone include the Middle Cambrian Scottsmore quartzite and various conglomeratic units in the Quebec group that

are of Middle(?) and Late(?) Cambrian, and of Early and Middle Ordovician age. Repeated brief subsidence and deposition followed quickly by rapid uplift and erosion, in the foreland belt and in adjoining areas immediately east of the foreland, probably explain these conglomerates (Cady, 1945, p. 538–539).

Quartz conglomerates, which occur principally within the eugeosynclinal zone and contain phenoclasts chiefly of vein quartz (Billings, 1956, p. 23; Currier and Jahns, 1941, p. 1497), were derived from terranes within the eugeosynclinal zone that had undergone tectonic consolidation, magmatic intrusion, and accompanying mineralization. The conglomeratic facies of the Lower Silurian Clough quartzite in northern New Hampshire contain phenoclasts of vein quartz that were eroded from the terrane in which the nearby Upper Ordovician Highlandcroft plutonic series was intruded (J. B. Thompson, Jr., 1958, oral communication) (Pl. 2). Other similar sources, most of which were first uplifted at about the time of the Taconic disturbance, probably existed in eastern areas of the eugeosynclinal zone. A. J. Boucot (1958, oral communication) has identified a local source of coarse debris in Somerset County in northwestern Maine that persisted during Silurian and Early Devonian time and to which he refers, in the paleogeographic sense, as "Somerset Island." This source, which he calls a "basement complex" is terrane that Marleau (1958, p. 3) has referred to as a pre-Silurian "complex" and the present author has tentatively assigned to the Middle Ordovician (Pl. 3).

The sources of the phenoclasts of some of the conglomerates in the eugeosynclinal zone and vicinity can only be inferred. The vein quartz cobbles and boulders in the conglomerates of the Middle Ordovician Umbrella Hill formation and the Silurian Shaw Mountain formation in Vermont are unlike the rocks of any exposed formations stratigraphically beneath them. They must have been derived from source terranes to the east, now buried beneath the Moretown, Northfield, or higher formations (Pl. 2). Westward thinning of the conglomerate of the Umbrella Hill formation, presumably away from an eastern source, is indicated by the absence of this conglomerate on the west limb of the Lowell Mountain anticline, although it is present on the east limb. Tectonic islands such as those that provided the phenoclasts in the Clough formation were probably the source. A similar source probably accounts for large isolated blocks in the Barton River formation (Irasburg conglomerate). These blocks reportedly include granite, andesite, diorite, diabase, serpentine, chlorite and sericite schists, sericitic marble, micaceous quartzite, crystalline limestone, black mudstone, and phyllite (Doll, 1951, p. 32–33; Richardson, 1929?, p. 108–109).

QUARTZ SANDSTONE: Quartz sandstone, which is the principal clastic constituent of the carbonate-quartzite assemblage, is derived chiefly from the direction of the craton. This is shown by quartz sandstone beds that occur in several different Cambrian and Ordovician stratigraphic units in the miogeosynclinal zone in west-central Vermont. These beds of sandstone thicken and become more abundant westward toward the craton (Cady, 1945, p. 534, 536, 546, 547) and pinch out to the east, commonly between beds of dolomite. Some of the quartz sandstones, notably the Lower Cambrian Gilman quartzite and the Middle Cambrian Scottsmore quartzite (=Rugg Brook formation), extend into the eugeosynclinal zone. The ultimate source of the quartz sands in these rocks was the siliceous Precambrian basement of the craton.

The source of quartz sands in the Lower Silurian Shaw Mountain, Peasley Pond, and Clough formations, which are in the miogeosynclinal and eugeosynclinal zones east of the Green Mountain anticlinorium is not so clear. Derivation from the Precambrian basement of the craton seems unlikely, inasmuch as this basement probably remained blanketed by Lower and Middle Ordovician carbonate rocks that were not breeched until long after the Early Silurian (Pl. 2). However, some of the quartz sands in the Lower Silurian rocks east of the anticlinorium probably came from land area to the west that was continuous with the craton but had not risen above sea level in the orthogeosynclinal belt until the time of the Taconic disturbance. F. F. Osborne (1957, written communication) reports that quartzite northeast of the Chaudiére River in Quebec, in the southeast limb of the Green Mountain anticlinorium (Gorman, 1954, p. 4), contains cross-bedding that indicates a source to the northwest. Quartz sand in this quartzite and in other quartzites in the east limb of the anticlinorium, including those of the Shaw Mountain and Peasley Pond formations, were possibly derived from Middle or Upper Ordovician rocks of the graywacke-shale assemblage, in the western part of the eugeosynclinal zone and in the exogeosyncline.

Much of the quartz sand, not only in the Lower Silurian quartzites, but also in quartz-

ites in Cambrian and Ordovician rocks beneath, was probably derived from within the eugeosynclinal zone. Islands in the eugeosynclinal zone are the probable source of the Clough quartzite, which occurs fairly far to the east, in the east limb of the Boundary Mountain anticlinorium. Islands that were the result either of prolonged failure of portions of the eugeosynclinal zone to subside or of volcanic accumulation are possible sources. The Middle Ordovician rocks of the eugeosynclinal zone, which include the graywacke-shale assemblage and rhyolitic volcanic rocks, contain the necessary quartz. The principal sources of quartz sand in the Clough quartzite, however, were probably tectonic islands that were raised in the eugeosynclinal zone and intruded by granitic igneous rocks at the time of the Taconic disturbance.

HISTORY OF SOURCE RELATIONSHIPS OF THE CLASTIC ROCKS: The author's interpretation of the history of the source relationships of the clastic rocks deposited in the orthogeosyncline may be summed up as follows:

The oldest clastic rocks known, which were deposited in Cambrian(?) time and were of the graywacke-shale assemblage, derived their sediments from mountainous terrain at the eastern margin of the craton. This terrain and adjacent submarine slopes to the east were reduced to low relief at the beginning of Early Cambrian time, in Vermont and nearby parts of Quebec, and before the Middle Ordovician in parts of Quebec north of the upper Yamaska River. With this reduction in relief the volume of sediments derived from the craton was reduced, and the carbonate-quartzite assemblage characteristic of the miogeosynclinal zone was established.

The sediments of most of the clastic rocks were derived from within the orthogeosyncline. Sources of limestone and dolomite boulders were provided by intermittent emergence of the miogeosynclinal zone and contiguous parts of the eugeosynclinal zone during the Cambrian and Early and Middle Ordovician time. Tectonic islands, on which pre-Silurian rocks consolidated by metamorphism and igneous intrusion were exposed, rose in the eastern part of the eugeosynclinal zone. The rise of the islands began at least as early as Middle Ordovician time and probably continued through the Silurian and Early Devonian. Clastic sediments eroded from the tectonic islands and also from archipelagic volcanic accumulations during Cambrian and Early and Middle Ordovician time formed thick deposits of the graywacke-shale assemblage in the western part of the eugeosynclinal zone. Tongues of these sediments overlap unconformably westward in the miogeosynclinal zone (Pl. 2); the most extensive tongue is of Middle Ordovician age.

Most of the eugeosynclinal zone was emergent during the Late Ordovician, and the uplifted terrain supplied clastic sediments, chiefly to shales that were deposited in the exogeosyncline to the west. Deposition in the exogeosyncline was brought to a close by an episode of general emergence at the time of the Taconic disturbance. At this time, consolidation of the exogeosyncline and underlying rocks of the miogeosynclinal zone, with the craton, began. The western margin of the miogeosynclinal zone moved eastward, probably beyond the eastern limits of the foreland, and the quasicratonic zone left behind it to the west (Pl. 2) became the source of some of the sediments which, in the Early Silurian, entered the carbonate-quartzite assemblages of the newly located miogeosynclinal zone.

A linear tectonic island or chain of islands near what was to become the site of the Boundary Mountain anticlinorium was a source of sediments within the eugeosynclinal zone during the Silurian and Early Devonian. The sediments of Silurian quartzite and conglomerate and Silurian and Lower Devonian graywacke and shale that were deposited in the eastern part of the eugeosynclinal zone came from this and probably other uplifts that were to the east. Quartz sands in Lower Silurian rocks were derived from surfaces of relatively low relief developed on these islands in an episode of stillstand that followed the Taconic disturbance.

Stratigraphic Breaks

GENERAL STATEMENT: The Taconic disturbance, the Acadian and Appalachian orogenies, and a few lesser diastrophic episodes are recorded by stratigraphic breaks in New England, Quebec, and the Maritime Provinces. Although the Acadian and Appalachian episodes are not recorded by unconformities within northern Vermont and contiguous parts of southern Quebec they are discussed here because the related unconformities, especially that of the Acadian, probably extended into this region before erosion.

The stratigraphic breaks reflected by some unconformities may not be so great as first appearances seem to indicate. Rarely can it be shown, particularly in the eugeosynclinal zone,

that whole formations are missing beneath unconformities. In some places, facies differences in underlying rocks may produce the effect of stratigraphic discordance of overlying beds. Some sections, east of the axis of the Green Mountain anticlinorium in Vermont, are without internal stratigraphic discordance or apparent break from Cambrian(?) through Lower Devonian rocks.

The seeming conflict between discontinuous and apparently continuous sections in the eugeosynclinal zone can be resolved if the zone be considered one of predominant subsidence during Cambrian(?) to Devonian time, but in which locally, in Middle Orodovician time and more generally in Late Ordovician-Early Silurian time, stability prevailed or gave way to uplift. Very pure quartzite, which is the most common rock of the Clough formation, suggests a relatively short episode of stillstand in the eastern part of the eugeosynclinal zone at the close of the Late Ordovician and Early Silurian episode. Separate interbeds of pure quartzite in the Ottauquechee formation seem to indicate that several very brief episodes of stillstand occurred in the eugeosynclinal zone during the Middle Cambrian. Conglomerates indicate uplift and discontinuous deposition at the sources of the phenoclasts though not necessarily where they were deposited.

TACONIC DISTURBANCE: The unconformity that marks the Taconic disturbance is preserved between Middle Ordovician and Lower Silurian strata in areas east of the axis of the Green Mountain anticlinorium. Basal conglomerate and associated quartzite of Early Silurian age, including those of the Peasley Pond, Lake Aylmer, Shaw Mountain, and Clough formations succeed the unconformity.

The age of the Taconic break is indicated by the stratigraphic relationships of the Middle Silurian Fitch formation in the east limb of the Boundary Mountain anticlinorium in New Hampshire (projected into section B-B′, Pl. 2). The Fitch unconformably overlies one of the plutons of the Highlandcroft plutonic series (Billings, 1937, p. 500, 518; 1956, p. 48) and contains boulders of the Highlandcroft, 1 mile northwest of Littleton, New Hampshire. A time interval at least long enough to have allowed for denudational unroofing of the plutons is necessary. The radiometric age of these plutons is 385 ± 27 million years (Larsen method), or probably Late Ordovician (Lyons and others, 1957, p. 529–530); therefore, the stratigraphic gap in the vicinity of the plutons probably includes much or all of the Upper Ordovician as well as the Lower Silurian. The stratigraphic break is probably even greater in the section on the west limb of the Boundary Mountain anticlinorium (east limb of the Connecticut Valley-Gaspe synclinorium), where the Fitch and also the Clough are absent (Pl. 3).

Billings (1937, p. 518) points out that the proof of the unconformity is based primarily on areal mapping and that the Fitch formation may overlie any one of four formations. Commonly, the Fitch rests on the Clough quartzite, with which it is conformable, but it also overlies various formations of Ordovician age that are older than the Clough. Billings ascribes the discordance of formations below the Clough to gentle folding that correlated with the Taconic disturbance.

A fairly large stratigraphic break that allowed time for some uplift and erosion is suggested by relationships in the southeast limb of the Green Mountain anticlinorium, northeast of the Chaudière River (Fig. 1) near St. Luc, Quebec. Gorman (1954, p. 4) has described boulder conglomerate and quartzite at this locality that look much like the Clough quartzite of New Hampshire. The St. Luc rocks, which are probably of Silurian age, apparently overlap unconformably northwestward on the lowest strata of the Middle Ordovician Beauceville formation.

Ambrose (1942; 1957, p. 164, 169–170) reports that the Peasley Pond conglomerate cuts across folds and a fault in the underlying Middle Ordovician rocks near Lake Memphremagog. This relationship is not clear because critical contacts are covered by surficial deposits, although it is possibly suggested by the general distribution of formations in the vicinity.

The unconformity between Middle Ordovician and Lower Silurian strata in the Lake Memphremagog synclines and in the west limb of the Connecticut Valley-Gaspe synclinorium is at the base of the overlapping miogeosynclinal zone (Pl. 2). This unconformity probably extended west across the Green Mountain anticlinorium, before the latter was eroded, and joined an unconformity at the summit of the Upper Ordovician rocks contained in the exogeosyncline. The unconformities between Ordovician and Silurian rocks east of the axis of the Connecticut Valley-Gaspe synclinorium are within the offlapping eugosynclinal zone.

ACADIAN AND APPALACHIAN OROGENIES: The principal unconformities in New England and the Maritime Provinces are those that are immediately beneath secondary geosynclines. The Moat volcanics of late Permian(?) age

(Lyons and others, 1957, p. 538–542), which are contained in an epieugeosyncline, overlie an angular unconformity that bevels folded and regionally metamorphosed rocks of the Lower Devonian Littleton formation in the White Mountains of New Hampshire (Williams and Billings, 1938, p. 1025–1026). This unconformity, which marks the Acadian orogeny and was the main episode of folding in the New England-Maritime-Quebec region (Billings, 1948, p. 52), is the upper limit of the eugeosynclinal zone. It coincides with a change from sections that contain various amounts of predominantly mafic marine volcanic rocks (Billings, 1937, p. 487–495) to sections that contain chiefly felsic flows, tuffs, and breccias (Billings, 1956, p. 36). The rocks in the latter sections are presumably all terrestrial deposits and, unlike the orogenically consolidated rocks beneath, they have not been metamorphosed.

The original westerly extent of this unconformity depends chiefly how far to the west the rocks of the epieugeosyncline extended before they were eroded. The Green Mountain anticlinorium, which includes the Lake Memphremagog synclines in which the Silurian and Devonian rocks are folded, is mainly a product of Acadian orogeny. Therefore, the synclinoria that merge with and flank the anticlinorium on the west, as well as related thrust faults in the foreland area (Cady, 1945, p. 577–578), are probably also Acadian.

The Upper Triassic Newark group, or its correlatives, which are contained in several taphrogeosynclines, unconformably overlie rocks of the eugeosynclinal zone and of epieugeosynclines. The unconformities between the rocks of the taphrogeosynclines and underlying rocks in the epieugeosynclines, the youngest of which are late Permian or younger, record the Appalachian orogeny. This orogeny was most effective in the southeastern parts of New England and adjacent areas. The unconformities coincide with a change in the composition of the volcanic rocks from predominantly felsic to mafic.

MINOR STRATIGRAPHIC BREAKS: The Umbrella Hill formation (Albee, 1957; 1957, Ph.D. thesis, Harvard Univ., p. 45–48) of north-central Vermont is composed chiefly of a basal conglomerate that is evidence of a stratigraphic break, probably of Middle Ordovician age. This break is within a section in the eugeosynclinal zone (Pl. 2). The conglomerate crops out in a narrow belt, commonly less than 500 feet wide and about 20 miles long, between the Stowe and Moretown formations in the east limb of the Lowell Mountain anticline. North and south of the ends of this belt the Stowe grades upward into the Moretown by interlamination of the strata; therefore, the stratigraphic break appears to be local. This seems true despite the fact that the pebble and cobble content suggests a large source terrain and that an apparent angular unconformity has been discovered between an isolated mass of the conglomerate and the underlying Stowe formation (Albee, 1957; 1957, Ph.D. thesis, Harvard Univ., p. 46). Various carbonaceous and noncarbonaceous schist and greenstone or amphibolite members of the Stowe formation are overlain by the conglomerate, which also suggests unconformable discordance, although lateral facies variations in the Stowe may be equally responsible for such a pattern.

Basal conglomerates, which are in about the same stratigraphic position as the conglomerates of the Umbrella Hill formation, have been reported at the bottom of the Beauceville formation in Quebec, chiefly north and west of Lake Memphremagog (Cooke, 1950, p. 34–36). The author found, however, that most of the conglomerate in the Beauceville is intraformational, and the underlying Mansonville formation grades into the Beauceville in the same way that the Stowe formation grades upward into the Moretown, north and south of the Umbrella Hill formation in Vermont.

West of the Green Mountain anticlinorium, the basal Mystic conglomerate is beneath the Stanbridge slate, which correlates with the upper part of the Beauceville formation. The Stanbridge (=Morses Line) correlates with several formations comprised of shale and related rocks of Middle Ordovician age that regionally overlap older formations of various ages on the foreland (Cady, 1945, p. 560). The regional unconformity produced is apparently due to uplift of the miogeosynclinal zone and the margin of the craton. The effects of this uplift possibly extend eastward into the eugeosynclinal zone and include some of the conglomerates in the Beauceville formation.

Some unconformable relationships that have been reported between stratigraphic units in the eugeosynclinal zone are questionable. Field evidence (Cooke, 1950, p. 34, 40, 42, 57–61, 76–79, 84–86, 88–91) for angular unconformities in the Cambrian to Devonian section in the vicinity of Lake Memphremagog is elusive. Contacts that might furnish proof of critical relationships are not exposed, or where exposed the critical angularity is between bedding in one bed and transecting cleavage in a subjacent bed. Pre-Middle Ordovician meta-

morphism, as well as folding, intrusion, and veining reported at a few localities in Quebec (Cooke, 1955; Riordon, 1954, p. 6; 1957), can possibly be interpreted as effects of subsidence to deeper zones in which pressure and temperature were greater; in the absence of angular unconformities, they do not necessarily indicate orogeny.

Magmatic Features

GENERAL STATEMENT: Magmatic features reflect changes in the mobility of the orthogeosynclinal belt from comparatively plastic conditions during subsidence and orogeny to a brittle state following orogeny and consolidation with the craton. The intrusive rocks emplaced during subsidence and also during orogeny are predominantly concordant or nearly concordant, presumably with zones of primary weakness parallel to the bedding of the stratified rocks. The contacts of intrusive rocks emplaced in consolidated terranes cut across the stratification along secondary tension joints formed after orogeny. The climax of mobility, near the peak of orogeny, is marked by thick concordant plutons. Mafic volcanic rocks, which were extruded both early and late in the diastrophic history of the orthogeosynclinal belt, probably reflect conditions of tension in the bottom of the deeply subsiding eugeosynclinal zone, as well as in the consolidated postorogenic terrane.

INITIAL MAGMA: Initial magmatic features include chiefly mafic and ultramafic igneous rocks—volcanic rocks that are interbedded with the stratified rocks of sedimentary origin in the orthogeosyncline and intrusive rocks that are emplaced in the volcanic and sedimentary accumulations. These features are characteristic of the eugeosynclinal zone.

The distribution of the volcanic rocks, which are predominantly mafic though locally felsic, has already been pointed out in the discussion of the graywacke-shale assemblage, with which they are interbedded. They offlap eastward away from the craton. Thus, those that are farthest west, in the west limb of the Green Mountain anticlinorium, are of Cambrian(?) age, whereas the westernmost volcanic rocks that are of Early Devonian age are far to the east in the east limb of the Connecticut Valley-Gaspe synclinorium.

Mafic and ultramafic intrusive rocks include chiefly sills and nearly concordant dikes and plutons of diabase, gabbro, dunite, peridotite, and serpentinite that have about the same distribution as the volcanic rocks. The mafic and ultramafic intrusive rocks that occur in the Green Mountain anticlinorium are probably of Late Ordovician age (Clark and Fairbairn, 1936, p. 16), whereas those to the east in the vicinity of the Boundary Mountain anticlinorium are as young as the Middle Devonian (Billings and White, 1950; Marleau, 1958, D.Sc. thesis, Laval Univ., p. 117, 120).

SYNOROGENIC MAGMA: Plutonic rocks that are of intermediate to felsic composition, notably those of the Highlandcroft and New Hampshire plutonic series (Billings, 1956, p. 46–48, 53–65), were intruded in the eastern part of the eugeosynclinal zone at about the times of the Taconic disturbance and the Acadian orogeny, respectively (Billings, 1956, p. 53, 108, 146, 150, 151–152; Eric and Dennis, 1958, p. 25–26; Lyons and others, 1957, p. 544).

The plutons of the Highlandcroft are irregular bodies (Billings, 1956, p. 122) that are emplaced in Middle Ordovician bedded rocks that are exposed near the axis of the Boundary Mountain anticlinorium (Billings, 1955; Eric and Dennis, 1958, Pl. 1). These plutons signify only temporary lapse of the eugeosynclinal regime at the time of the Taconic disturbance. This is shown by the recurrence in the Lower Devonian rocks on both limbs of the Boundary Mountain anticlinorium of initial magmatic features, notably mafic pillow lavas.

The plutons of the New Hampshire plutonic series include many large and nearly concordant bodies (Billings, 1956, p. 125–129) that are emplaced principally in the Silurian and Devonian bedded rocks on the eastern flank of the Boundary Mountain anticlinorium (Billings, 1955) and in the Connecticut Valley-Gaspe synclinorium (White and Billings, 1951, p. 662, 665–667). The New Hampshire plutonic series, which was intruded chiefly after the climax of folding and metamorphism during the Acadian orogeny (Billings, 1956, p. 146–147), marks the close of the orthogeosynclinal regime.

SUBSEQUENT MAGMA: Subsequent magmatic features include chiefly intermediate to felsic igneous rocks—volcanic rocks that are commonly interstratified with other bedded rocks in secondary geosynclines and intrusive rocks that invade the secondary geosynclines. Subsequent magmatic features are particularly characteristic of the epieugeosynclines, although this is not apparent in the northwestern part of the New England-Quebec-Maritime region where the volcanic rocks contained in the epieugeosynclines are almost entirely eroded, and all that remains are the stumps of intrusive bodies beneath the secondary geosynclines.

The volcanic rocks of subsequent magmatic origin include the Moat volcanics of the White Mountain plutonic-volcanic series (Billings, 1956, p. 35–37), which unconformably overlie Lower Devonian rocks that were folded and metamorphosed during the Acadian orogeny (Williams and Billings, 1938, p. 1025–1026). The intrusive rocks are characteristically the discordant plutons of the White Mountains (Billings, 1956, p. 69–88, 129–135) and the Monteregian plutonic series in Quebec (Dresser and Denis, 1944, p. 455–482), as well as various correlated mafic dike rocks (Billings, 1956, p. 86; Cooke, 1950, p. 105–110; Kemp and Marsters, 1893). The White Mountain and Monteregian series are of Permian age (Lyons and others, 1957, p. 538–542), and both are alkalic; therefore, they are probably comagmatic. Plutons of about this age and composition are distributed from Montreal, Quebec, to southeastern New England in a great arc that extends from the craton across the orthogeosynclinal belt and marks the spread of quasicratonic conditions eastward to the Atlantic Ocean.

FINAL MAGMA: Mafic volcanic and hypabyssal rocks of Late Triassic age that are contained in the taphrogeosynclines in southern New England and connected regions appeared after the Appalachian orogeny. Their appearance marked the close of geosynclinal history and signified that the geosynclinal belt had become a completely consolidated extension of the craton.

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Supporting Information for Stratigraphic Relationships Shown on Correlation Table (Pl. 3)

Several of the stratigraphic units shown on Plate 3 need supporting information in addition to that supplied in the publications cited at the bottom of the table. This information is annotated numerically below with reference to key numbers inserted in the correlation table. Geologically older units, beginning with those west of the Green Mountain anticlinorium, are discussed first. Items in the annotations that are not documented are observations or conclusions of the author.

1. The Call Mill slate (Clark, 1936b, p. 138–140), which has not hitherto been reported in Vermont, crops out 3.5 miles S. 63° W. of the village of Bakersfield, Vermont, in the east limb of the Enosburg Falls anticline. Here it is about 20 feet thick and is in contact with vesicular lavas of the Tibbit Hill volcanics to the west and stratigraphically beneath, and it grades stratigraphically upward into schistose facies of the Pinnacle formation to the east.

2. A section that contains interbeds of graywacke like that of the Pinnacle formation, a foot or more thick, but is predominantly of quartz-sericite-chlorite schist occurs in the Cambridge syncline in Vermont. Booth (1950, Pl. 1) informally names this terrane the "Jay Peak series," a designation that will probably be revised or replaced as mapping continues. Abundant lenticular segregations of granular white quartz are distinctive features of the schist. The schist increases in abundance eastward from the Enosburg Falls anticline with a corresponding decrease in the graywacke. Greenstone in this terrane at various places east of the axis of the Cambridge syncline is possibly an eastern extension of the Tibbit Hill volcanics of the Enosburg Falls anticline. Such a relationship suggests that the Tibbit Hill of the Enosburg Falls anticline is a member within rather than entirely below the Pinnacle formation. Osberg (1955, oral communication) reports such a relationship at the axis of this anticline a few miles north of the Lamoille River.

Similar rocks in Quebec, in a comparable stratigraphic position, are termed the "Bonsecours" formation by P. H. Osberg (1959, written communication). This term is used informally here for convenience in discussion.

The quartz-sericite-chlorite schist of the Cambridge syncline is similar to schistose facies of the West Sutton formation of the northern extension of the Enosburg Falls anticline near (Booth, 1950, p. 1147) and north of the international boundary and of the Gilman formation of this anticline in Quebec. Clark's description of these formations in Quebec (1936b, p. 143–146) applies to little metamorphosed rocks. The schist terrane of the Cambridge syncline therefore appears to include rocks synchronous with the West Sutton and Gilman formations, and it is also a facies that expands downward to the east, east of the Enosburg Falls anticline, at the expense of the various facies of the underlying White Brook dolomite, Pinnacle formation, Call Mill slate, and possibly also the Tibbit Hill volcanics.

3. The upper strata of the "Jay Peak series" of Booth (1950, Pl. 1) in the Cambridge syncline in Vermont are possibly younger than the upper beds of the Gilman quartzite of the Enosburg Falls anticline. Before the crest of the anticline was eroded these upper rocks probably extended westward and graded laterally into the Dunham dolomite and possibly the Parker slate.

The eastern limits of the quartz-sericite-chlorite schist terrane ("Jay Peak series") in Vermont have not been defined. To the east the schist becomes locally quartzitic, and interbeds of carbonaceous quartz-sericite-chlorite schist and carbonaceous quartzite appear. Albite porphyroblasts are a common feature of the schists of the latter association, and in some beds the albite is abundant enough to form a gneiss. This assemblage of carbonaceous as well as noncarbonaceous rocks makes up the Camels Hump group of the axial anticline of the Green Mountain anticlinorium. The top of the Camels Hump group is probably as high stratigraphically as the Parker slate.

4. In the west limb of the St. Albans synclinorium in Quebec, the Dunham dolomite remains unreported in a terrane of variously designated Cambrian rocks almost entirely dolomite (Clark, 1934, p. 7, 8; Ells, 1896, p. 35; Logan, 1863, p. 284–285, 851–852). Clark (1934, Fig. 3, p. 7) refers all the dolomite to the "Milton dolomite," that is, to strata higher than the Dunham. However, Shaw (1958, Pl. 1) has recently mapped the Dunham north to the international boundary—and it therefore presumably extends north of the boundary into Quebec—immediately east of and above the Champlain thrust (Pl. 1). The upper limits of the Dunham of the west limb of the St. Albans synclinorium in Quebec are uncertain inasmuch as the Parker slate, which caps the Dunham in Vermont, does not extend into Quebec. A fairly distinctive formation, the Saxe Brook dolomite, which probably overlies the Dunham of this terrane, is discussed in annotation 8.

A somewhat comparable dolomitic terrane, in northern areas of the west limb of the Enosburg Falls anticline in Vermont, contains rocks that are probably equivalent to the Dunham dolomite, Parker slate, and Rugg Brook formation. Rarity or absence in this terrane of the intervening slates, makes difficult the differentiation from the Dunham dolomite of dolomitic facies of the Parker and Rugg Brook. The Rugg Brook may be recognized, however, by its zones of dolomitic conglomerate and gray-weathered quartzitic sandstone crisscrossed by veins of white quartz.

5. The Rockledge conglomerate has been mapped in the west limb of the Enosburg Falls anticline (east limb of the St. Albans synclinorium) in Vermont chiefly on the basis of lithic features that are similar to those of rocks of the type locality of the Rockledge in the west limb of the St. Albans synclinorium (Shaw, 1958, p. 542). The boulders and

pebbles in this conglomerate are of limestone as at the type locality. The Rockledge is lithologically like the Mill River conglomerate; therefore, the conglomerate in eastern exposures that contains limestone phenoclasts, on the basis of lithic comparison, could be either Mill River or Rockledge, or both. The eastern exposures of the Rockledge have also been equated with the Rugg Brook formation (Booth, 1950, p. 1152–1153), but the Rugg Brook differs in that its phenoclasts are of dolomite rather than limestone. The conglomerates, in the west limb of the Enosburg Falls anticline, that contain phenoclasts of limestone lie on and intergrade with the Rugg Brook formation; thus, slates do not separate them from the Rugg Brook as in the west limb of the St. Albans synclinorium. Possibly the slates are missing and the conglomerates are indeed Rockledge, although it seems an equal possibility that these conglomerates are lower in the stratigraphic section than the type Rockledge.

6. Slates that are next west of the conglomerates that contain phenoclasts of limestone, in the west limb of the Enosburg Falls anticline (east limb of the St. Albans synclinorium) in Vermont, have been referred to the Hungerford slate, which overlies the Rockledge conglomerate on the west limb of the St. Albans synclinorium (Shaw, 1958, p. 547). If the conglomerates of the west limb of the Enosburg Falls anticline are stratigraphically lower than the Rockledge (*see* annotation 5) the slates to the west of the conglomerates possibly correlate with both the Skeels Corners slate and the Hungerford slate on the west limb of the St. Albans synclinorium. Should this be the case the term Sweetsburg slate, as used in the Enosburg Falls anticline in Quebec, seems more appropriate. These slates are lithologically more like the Skeels Corners or Hungerford than like the less siliceous Morses Line slate at the axis of the St. Albans synclinorium.

7. The extent of the Scottsmore quartzite and the Sweetsburg slate in the Cambridge syncline in Vermont and nearby areas of Quebec is not clear, pending completion of detailed geologic mapping. These formations apparently overlie rocks referred to in annotations 2 and 3, which are assigned to the "Jay Peak" series (Booth, 1950, Pl. 1). T. H. Clark (1951, oral communication) has shown the author rocks that he considers Sweetsburg, on the west limb of the Cambridge syncline near Sutton, Quebec. Christman (1959, Pl. 1) maps Ottauquechee (=Sweetsburg) in the vicnity of Cambridge, Vermont. P. H. Osberg (1958, written communication), maps Sweetsburg in the northern extension of the Cambridge syncline between Sutton and Richmond, Quebec (Fig. 1; Pl. 1).

8. The Saxe Brook dolomite of the west limb of the St. Albans synclinorium in Quebec has, like the Dunham beneath it (*see* annotation 4), remained unreported in a terrane of various Cambrian rocks referred by Clark (1934, Fig. 3, p. 7) to the "Milton dolomite." The Saxe Brook of this terrane to the south in Vermont (Shaw, 1958, Pl. 1) occupies the interval between the Parker slate and the Hungerford slate. The Saxe Brook is differentiated from the underlying Dunham dolomite and the overlying Hungerford slate at the international boundary (Shaw, 1958, Pl. 1) and presumably is a mappable unit north of the boundary.

9. The Gorge formation crops out in the west limb of the St. Albans synclinorium in Quebec, notably at Rosenberg, Quebec, where it plunges north, at the axis of a minor anticline, beneath the slaty limestone of the Highgate formation. It is apparently cut out to the west at the Champlain overthrust. The Gorge is the uppermost formation in the terrane that Clark (1934, Fig. 3, p. 7, 8) referred to the "Milton dolomite."

10. The "Corliss" conglomerate of the St. Albans synclinorium in Vermont crops out in isolated exposures within the terrane of the Morses Line slate and is considered by Shaw (1951, p. 98; 1958, p. 553) to be a part of the Morses Line. The structural relationships of these masses of conglomerate are not clear; possibly they are near the base of the Morses Line slate in a position comparable to that of the Mystic conglomerate beneath the Stanbridge slate. If this is true, a correlation of the "Corliss" with the Mystic seems likely.

11. The Morses Line slate of the St. Albans synclinorium has been mapped north in Vermont to the international boundary by Shaw (1958, Pl. 1). The slates immediately north of the boundary in the St. Albans synclinorium have been referred to the "Georgia slate" (Clark, 1934, p. 7, Fig. 1, p. 8), in the same way as were those south of the boundary before Shaw's work (Keith, 1932, p. 378–379; Schuchert, 1937, p. 1052). It is now logical to extend the use of Shaw's name—Morses Line slate—to the contiguous rocks in Quebec. The Morses Line, so extended, merges northward with the lithologically similar Stanbridge slate of the Philipsburg slice, and a correlation of the Morses Line with the Stanbridge is inferred. The two slates join where the Champlain overthrust, between the St. Albans synclinorium and the Philipsburg slice to the west, dies out northward near Rosenberg, Quebec.

12. The Clarendon Springs dolomite, Shelburne limestone (=Shelburne marble), Cutting dolomite, and Bascom formation of the Champlain-Richelieu foreland in Vermont were early referred to as "Calciferous A," "B," "C," and "D," respectively (Brainerd and Seely, 1890a, p. 506–508; 1890b, p. 10–22), and later, with change of terminology, became "Little Falls dolomite" and "Beekmantown B," "C," and "D" (Clarke and Schuchert, 1899; Clarke, 1903, p. 16 and chart). The terms Clarendon Springs, Shelburne, Cutting, and Bascom are extensions of the terminology used in the Hinesburg and Middlebury synclinoria of west-central Vermont (Cady, 1945, p. 536–545) and are applied to similar rocks in the area west of the foreland thrust slices. The general correlation and lithic similarity of these rocks to units in the Philipsburg slice has long been recognized (Bradley, 1923, p. 331–333; Brainerd and Seely, 1890a, p. 508; 1890b, p. 23). If present on the Champlain-Richelieu foreland of

Quebec the Clarendon Springs, Shelburne, Cutting, and Bascom are probably beneath the bottoms of wells drilled to date. These formations crop out on the Champlain-Richelieu foreland of Vermont south of about latitude 44°45'N.

13. The Burchards member of the Chipman formation, which crops out south of Burlington in west-central Vermont, was formerly correlated with the middle Chazy Crown Point limestone (Brainerd, 1891, p. 299; Cady, 1945, p. 548–550) but is now known to be not only beneath the Crown Point limestone (Kay and Cady, 1947) but beneath the Chazy (Kay, 1950). Studies by E-an Zen and the author show that the Burchards is discontinuous laterally and is apparently a time equivalent chiefly of the lower part of the Beldens formation discussed in annotation 14 (Cady and Zen, in press). The Burchards member is lithically, faunally, and sequentially like the Solomons Corner formation of the Philipsburg slice (Clark and McGerrigle, 1944, p. 399; Logan, 1863, p. 844, division B4), and a correlation of the two is therefore shown on Plate 3. Characteristic of these two units—the Burchards and Solomons Corner—are dolomitic mottling, species of the gastropod *Maclurites*, and the general absence of beds of massive dolomite.

The Burchards is missing from the Highgate Springs-St. Dominique slice, possibly because it is cut out by the Highgate Springs-St. Dominique thrust rather than because it is absent from the section in northwestern Vermont. This explanation is suggested by the proximity of the Solomons Corner formation in nearby Quebec.

14. The Bridport dolomite and the Beldens formation are western and eastern facies, respectively, of contemporaneously deposited rocks (Cady and Zen, in press; Kay, 1950; 1958, p. 79; Kay, *in* Twenhofel and others, 1954, p. 263). The Bridport dolomite of the Champlain-Richelieu foreland in Vermont was early referred to as "Calciferous E" (Brainerd and Seely, 1890a, p. 507; 1890b, p. 14–20), which with change of terminology (Clarke and Schuchert, 1899) became "Beekmantown E." It is also known as the Providence Island dolomite (Erwin, 1957, p. 11, 63; Ulrich and Cooper, 1938, p. 26). The Beldens formation is composed of interbedded limestone and dolomite that were formerly assigned to the upper Chazy (Brainerd, 1891, p. 299; Cady, 1945, p. 548, 552) but are now known to be of the uppermost Beekmantown (Kay, 1958, p. 79).

The St. Armand and Corey formations of the Philipsburg slice (Clark and McGerrigle, 1944, p. 399–400; Logan, 1863, p. 844–845, divisions B5 through C1) are remarkably similar to the typical Beldens except that they contain less dolomite. Probably they make up a more easterly facies of the transition from the Bridport east to the Beldens referred to above. Like the Beldons and associated strata of west-central Vermont, the St. Armand and Corey were at first thought to correlate with the Chazy (Brainerd, 1891, p. 298–299).

15. The subsurface Beldens of the deep wells in the Champlain-Richelieu foreland in Quebec (Belyea, 1952, p. 11–14; Clark, 1955, p. 10) may reflect a northward as well as an eastward transition into the interbedded limestone and dolomite of the Beldons from the Bridport dolomite of the Champlain-Richelieu foreland in Vermont.

16. The "Sherbrooke" of the Stoke Mountain anticline in Quebec (Cooke, 1951) refers to several lithic types (Cooke, 1950, p. 49–56) in various stratigraphic units including Mansonville and Beauceville. Coarse conglomeratic quartzite and graywacke that have been referred to the "Sherbrooke" are identical with rocks in the Mansonville formation. Such a correlation was apparently implied by Ambrose between his Unit 1 in the Stanstead map area, which covers the south end of the Stoke Mountain anticline, and his Unit 2 in the Mansonville map area (Ambrose, 1942; 1943). Volcanic rocks, both felsic and mafic, that have been referred to the "Sherbrooke," are interbedded chiefly in the Beauceville slate and are probably stratigraphically equivalent to the Bolton volcanic rocks of the vicinity of the Lake Memphremagog synclines. "Argillite" and "grit" that have been referred to as "Sherbrooke" (Cooke, 1957, p. 21–28) are interbedded in the St. Francis group of the Connecticut Valley-Gaspe synclinorium (Marleau, 1958, D.Sc. thesis, Laval Univ., p. 93–94; J. B. Thompson, Jr., 1957, oral communication).

17. The term "Caldwell" has been used for rocks stratigraphically beneath the Beauceville slate in Quebec (Cooke, 1937, p. 11–33; 1950, p. 19–28), down to and including the Sutton schist (p. 19–23), at the base of the exposed section in the Green Mountain axial anticline. It was originally used, however, in a more restricted sense, with reference to rocks between the Sutton and Beauceville (Dresser and Denis, 1944, p. 340; MacKay, 1921, p. 20–24, Map 1835) and stratigraphically about equivalent to the Mansonville. The "Caldwell" is not recognized here because of its ambiguous use and also because the word "Caldwell" apparently originated as a misspelling of the word "Colway," which is the name of a brook at the type locality (Dresser and Denis, 1944, p. 340, footnote). In the Stoke Mountain anticline the "Caldwell" (Cooke, 1950, p. 22) appears to be a predominantly quartzitic schist facies intermediate in coarseness and in position in the section between the "Sherbrooke" graywacke and conglomerate and slates referred to the Beauceville. The "Caldwell facies" of De Romer (1957, p. 5–6; 1958, p. 5–7) refers to the upper part of the Mansonville and most nearly approximates the original usage of the term "Caldwell."

18. The slates in Quebec that are referred to the Beauceville (Cooke, 1950, p. 37) grade downward and laterally into conglomerate, grit, and quartzite referred to the "Sherbrooke" and "Caldwell," in and adjacent to the Lake Memphremagog synclines. The boundary between the Beauceville and underlying rocks is not sharp; there is an almost imperceptible gradation by interlamination of the slates with the coarser clastic rocks, and boundary

mapped is undoubtedly at higher time-stratigraphic horizons in some places than in others. Conglomerates in the lower part of the Beauceville of the Stoke Mountain anticline and of the Lake Memphremagog synclines and vicinity are intraformational, and the boulders and pebbles are commonly of the grit and quartzite. Unconformities, some of them angular, have been reported between the Beauceville and underlying rocks or in the upper part of the "Caldwell," in areas northeast of Lake Memphremagog (Clark, 1937a, p. 35–36; Cooke, 1950, p. 40–42, 1955; Riordon, 1954, p. 6, 8; 1957). The unconformable relations reported by Cooke (1950, p. 40–42) in the area immediately north of Lake Memphremagog seem very doubtful.

19. The Moretown formation of the Lowell Mountain anticline and of the Lake Memphremagog synclines in Vermont contains more graywacke and quartzite than does the equivalent Beauceville in Quebec, although slate beds such as those typical of the Beauceville are common. Farther south in Vermont toward the type locality of the Moretown, south of the Winooski River, the slates are uncommon in some sections. Unlike the Beauceville of the Lake Memphremagog synclines and vicinity of Quebec, the Moretown is underlain by conglomerate, in the Umbrella Hill formation (Albee, 1957; 1957, Ph.D. thesis, Harvard Univ., p. 45–48), that extends north to within about 10 miles of the international boundary in the east limb of the Lowell Mountain anticline. The contact betaeen the Moretown and the underlying Stowe formation beyond the northern and southern limits of this conglomerate is gradational. The Cram Hill formation (Currier and Jahns, 1941, p. 1493–1496), which grades laterally into the upper part of the Moretown formation south of Montpelier, Vermont, is characterized by slate and metamorphosed volcanic rocks.

20. The Bolton mafic lavas of the Lake Memphremagog synclines in Quebec are, as Ambrose has shown (1942; 1949; 1957), interbedded and folded with the strata of the Beauceville slate. (*See* Fortier, 1945, p. 1.) These lavas extend down to or nearly to the base of the Beauceville in Quebec, but at the international boundary and south of the boundary in Vermont a considerable thickness of slate such as that in the Beauceville, as well as typical granulite, quartzite, and phyllite of the Moretown, lies beneath and also above the volcanic rocks. The rocks beneath the volcanic rocks in Vermont interfinger laterally northward across the international boundary into the lower part of the Bolton lavas in Quebec. Therefore, the Bolton volcanics in effect form a tongue that diminishes in thickness southward from Quebec into Vermont and gradationally overlaps the lower part of the Moretown formation.

The Bolton volcanic rocks and the volcanic rocks of the "Sherbrooke group" of Cooke (1950, p. 52–56) probably correlate with the Coleraine group of Riordon (1954, p. 6–7), in the Thetford Mines-Black Lake area (Fig. 1). They probably also correlate with volcanic rocks northeast of the Thetford Mines-Black Lake area that occur in both the upper "Caldwell" and in the Beauceville (Béland, 1952, p. 5, 6; 1954a, p. 4; 1954b, p. 3; 1957, p. 16–17, 20–21; Dresser and Denis, 1944, p. 425–427; Gorman, 1954, p. 2–3; MacKay, 1921, p. 20–26; Tolman, 1936, p. 6–7, 11–12).

21. The Ammonoosuc volcanics are known on the west limb of the Boundary Mountain anticlinorium in eastern Vermont south-southeast of St. Johnsbury (Eric and Dennis, 1958, Pl. 1) and in neighboring New Hampshire. Here they grade (*see* Billings, 1937, p. 476, Pl. 1) eastward into successively older parts of the Albee formation near the axis of the Gardner Mountain anticline, the axial anticline of the Boundary Mountain anticlinorium in this area.

To the north, near the northern tip of New Hampshire and in adjacent parts of Maine and Quebec, various correlations of the Ammonoosuc are suggested by lithologic comparisons. Correlation with the "Arnold River formation" of Marleau (1958, D.Sc. Thesis, Laval Univ., p. 32) is suggested by similar stratigraphic position immediately beneath known Silurian and Devonian rocks and by comparable structural position near the axis of the Boundary Mountain anticlinorium, as well as by lithologic comparisons. The "Arnold River formation" is a "complex" of metamorphosed sedimentary rocks exposed at the axis of the anticlinorium along the frontier between Quebec and Maine (A. L. Albee, 1957, oral communication; Marleau, 1958, Map 1217; 1958, D.Sc. Thesis, Laval Univ., p. 32). It is stratigraphically beneath the Seboomook formation which is tenatively assigned an Early Devonian age in Quebec (Marleau, 1958, D.Sc. Thesis, Laval Univ., p. 43–44). Rocks of the "Arnold River formation" that adjoin the Seboomook are lithologically similar to sedimentary interbeds in the Ammonoosuc. A tentative correlation of the Albee and Ammonoosuc with the "Arnold River" is shown in Plate 3.

Volcanic rocks similar to the Ammonoosuc volcanics of Ordovician age, but only in part correlative with them, crop out west of the Albee formation in northernmost New Hampshire. These volcanic rocks contain interbeds of metamorphosed sedimentary rocks (Billings, 1956, p. 14; Huntington, 1877, p. 47–50), and they are intruded by ultramafic rocks (Billings, 1955; 1956, p. 46). The presence of the ultramafic rocks has suggested correlation of these volcanic rocks with the Ammonoosuc, inasmuch as ultramafic rocks have been proven to intrude only Ordovician or older rocks in New England. Part of the volcanic and associated metasedimentary rocks are in a terrane that has been assigned to the Ordovician Orfordville formation (Billings, 1956, p. 14). It is difficult to distinguish the metamorphosed sedimentary rocks of this terrane from that of the Gile Mountain formation (Devonian), which is next west.

Thus, rather than being Ammonoosuc or Orfordville most of the volcanic and associated rocks in northernmost New Hampshire and adjacent Maine

are probably a southern extension of the Devonian Seboomook and Frontenac formations in immediately adjacent Quebec, with which they appear to be continuous (Marleau, 1958, D.Sc. Thesis, Laval Univ., Map 4). Both the Seboomook and the Frontenac, which overlies the Seboomook, are intruded by ultramafic rocks (p. 120). These rocks are a manifestation of ultramafic activity later than that shown by rocks in the belt of the Green Mountain anticlinorium, but, as in the latter belt also, it seems most significant that the distribution of the ultramafic rocks in space and time agrees with that of the volcanic rocks with which they are associated.

22. The St. Francis group or "series" of the Connecticut Valley-Gaspe synclinorium in Quebec, which is in general synonymous with the term "Compton formation," or locally "Tomifobia formation," is included in a terrane that was originally assigned to the Silurian and Devonian by Logan (1849, p. 57–58; 1863, p. 427–437) followed by Hunt (1854, p. 197–198) and Selwyn (1883, p. 1–2). Ells (1887, p. 7–14, 21–22; 1896, p. 41) and most subsequent authors (Ambrose, 1943; Clark, 1934, p. 11, 12; 1937b, p. 42–43; Cooke, 1950, p. 32; 1957, p. 19–21; Faessler, 1939, p. 8; McGerrigle, 1935, p. 73) have assigned St. Francis, Compton, or Tomifobia to the Ordovician. However, rocks of the St. Juste group, which is the northeastern extension of the St. Francis in Quebec, have very recently been placed in the Silurian(?) and Devonian (Béland, 1952, p. 6; Gorman, 1954, p. 3–4; 1955, p. 4–5). Correlation with the fossiliferous Devonian "Famine series" (MacKay, 1921, p. 31–32) is apparently the principal basis for this assignment. The Compton is now specifically referred to the Silurian or Devonian (Marleau, 1958, p. 3–4), and the St. Francis group is assigned a Silurian and Devonian age on the current geologic map of Quebec (Quebec Department of Mines, 1957).

With a single exception, diagnostic fossils are not known in rocks that to date have been referred explicitly (Clark, 1937b; Cooke, 1950, p. 29–32; Faessler, 1939, p. 10–11; McGerrigle, 1935, p. 71–73) to the St. Francis, Compton, or Tomifobia. Authors, beginning with Logan, who have placed rocks now referred to these formations in the Silurian and Devonian, have depended upon correlations involving rocks that are not mapped as St. Francis, Compton, or Tomifobia. The original assignment of the St. Francis terrane to the Ordovician apparently arose from Ells' (1887, p. 21) conclusion that the rocks of this terrane are unconformable beneath the Silurian. This in turn may probably be attributed to Ells' misinterpretation of Logan's (1849, p. 47–48) stratigraphic assignment of the rocks beneath the unconformity. Logan was referring to rocks now called Beauceville in the west limb of the Stoke Mountain anticline, whereas Ells apparently thought he meant rocks now called St. Francis in the east limb (west limb of the Connecticut Valley-Gaspe synclinorium).

The assignment to the Ordovician was based on fossil collections reported from two localities. One is northeast of Lake St. Francis, in rocks originally included in the Beauceville (Cooke, 1933, p. 45) and later shifted to the St. Francis without explanation (Clark, 1937b, p. 42–43). The other locality, which is near the international boundary east of Lake Memphremagog (Tomifobia formation), yielded fossils (Clark, 1934, p. 12) that were inadequately described in the literature and, therefore, have been of little use as a basis for correlation. Upon re-examination by D. J. McLaren and L. M. Cumming of the Geological Survey of Canada the age of the latter fossils was found indeterminate (see discussion of geologic age).

The terrane that Logan, Hunt, and Selwyn called Silurian and Devonian included not only the present St. Francis, Compton, or Tomifobia, but also critical fossiliferous rocks of post-Ordovician age that have since been differentiated as the Lake Aylmer group or "series" (Burton, 1931, p. 118–120; Clark, 1937c; Cooke, 1950, p. 71–76) and the "Famine series" (Clark, 1923; MacKay, 1921, p. 31–32; Tolman, 1936, p. 16). The St. Francis or its equivalents are mapped by most authors (Cooke, 1938, 1951; Cooke and Clark, 1938a, 1938b) in fault contact with the Lake Aylmer rocks, so that the St. Francis appears to be structurally isolated from the rocks that furnished Logan, Hunt, and Selwyn one of the principal clues as to its age. If the faults do exist correlation is difficult, but if they do not exist, as seems likely, Logan's correlation must be reconsidered.

These faults and their extensions (Ambrose, 1943; Clark, 1934, p. 4; Doll, 1951, p. 52–55) are, with few exceptions, mapped in areas of surficial deposits, where their existence is only conjectural, as for example in the belt at the east foot of the Lowell Mountains in Vermont and the Stoke Mountains in Quebec, where the alluvial deposits along the St. Francis River and other streams cover large areas. South of Marbleton, a fault is mapped (Cooke, 1951) in an area of fairly abundant outcrops, between the Lake Aylmer group to the west and the St. Francis group to the east. The author observed here a gradation from the silty, brown-crusted limestone typical of the St. Francis, on the east, into the slightly cleaner limestones of the Lake Aylmer to the west. This is apparently the relationship noted by Laverdière (1936, p. 35–38), who did not map a fault here. Only locally in the Lake Aylmer are there clean limestone masses that seem strikingly different from the St. Francis. Some of the slaty limestone of the Lake Aylmer formation resembles the Glenbrooke shale of the Lake Memphremagog synclines, with which it probably correlates. Ambrose (1943) remarked on the resemblance of the St. Francis to "Devonian" rocks near Lake Memphremagog, although he did not intend a correlation. C. G. Doll and the author recently examined the Lake Aylmer group 3 miles southeast of Weedon, Quebec, where it adjoins the "Sherbrooke group" (Cooke, 1951), and find that the rock near the contact is similar to the Northfield slate, and that the rock next west is like Doll's Ayers Cliff formation, both of which are

in the west limb of the Connecticut Valley-Gaspe synclinorium in Vermont.

Briefly then, the rocks now referred to the St. Francis group were, more than 100 years ago, placed in the Silurian and Devonian on fairly good stratigraphic evidence. Later, they were assigned to the Ordovician, apparently through confusion with rocks now included in the Beauceville. The St. Francis group correlates with formations to the northeast, in the Connecticut Valley-Gaspe synclinorium, that have recently been assigned to the Silurian and Devonian on essentially the same evidence as employed a century ago. This correlation is supported by the gradational relationships between the St. Francis and the fossiliferous Lake Aylmer group.

23. The Gile Mountain formation of northeastern Vermont is believed by the author to be in the section stratigraphically above the Ammonoosuc volcanic rocks in the west limb of the Boundary Mountain anticlinorium. Billings (1956, p. 94, 98) suggests that the Gile Mountain formation may be in a higher stratigraphic position than the Ammonoosuc, although an earlier interpretation (Billings, 1955) had tentatively placed the Gile Mountain in a stratigraphic position well beneath the Ammonoosuc. Detailed mapping now in progress in northeastern Vermont and adjacent New Hampshire should furnish the additional facts necessary to show the true relationships.

In northernmost Vermont the Gile Mountain is continuous westward in the Connecticut Valley-Gaspe synclinorium with identical slate, graywacke, and quartzite referred to the Westmore formation (Doll, 1951, p. 33–37), although farther south in Vermont belts of silty limestone are included in such terrane. Where this occurs the belts of limestone are variously referred to as the Waits River formation (Dennis, 1956, p. 16–19; White and Jahns, 1950, p. 187–188), or Barton River formation (Doll, 1951, p. 25–32), or both (Murthy, 1957, p. 22–24; 1958). Similar rocks with limestone interbeds crop out east of the Gile Mountain formation in northern New Hampshire, where they have been referred to the Waits River formation (Billings, 1955; 1956, p. 6–8).

It is not known to what extent the belted areal distribution of the carbonate rocks in the Connecticut Valley-Gaspe synclinorium reflects lateral change of facies with interlamination, from abundant limestone to the west to little limestone to the east (Murthy, 1957, p. 46, Pl. 3), and how much of it reflects folding and dissection of an alternating stratigraphic section with abundant limestone at the base but with none at the top (Dennis, 1956, p. 16–22). The latter interpretation is suggested on the restored sections (Pl. 2). There seems little doubt that the uncertainties are due to complex relationships of both facies and structure (Eric and Dennis, 1958, p. 58–61). (*See* Dennis, 1959; Murthy, 1959a; 1959b; White, 1959.)

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PALEOCURRENT STRUCTURES AND PALEOGEOGRAPHY OF THE ULTRAHELVETIC FLYSCH BASINS, SWITZERLAND

By K. Jinghwa Hsu

Abstract

The Ultrahelvetic Flysch of the Swiss Prealps ranges from Late Cretaceous to Eocene in age. It occurs as isolated outcrops which are known by various local names—Wäggital Flysch in the Glarus Alps, Schlieren Flysch in central Switzerland, Leissigen Flysch in the Internal Prealps southwest of Lake Thun, and Gurnigel Flysch in the Bernese External Prealps. These Flysch formations are characterized by thick sequences of interbedded sandstone, conglomerate, shale, and silty limestone. The sandstone of the Flysch exhibits sedimentary structures typical of turbidity-current deposition—*e.g.*, graded bedding, flute casts, groove casts. About 400 measurements of the directional sedimentary structures were made from the various Flysch formations. Paleocurrent information thus obtained was interpreted within the framework provided by the stratigraphic and tectonic studies of alpine geologists.

The major conclusions of this paper are:

(1) The Ultrahelvetic Flysch basins of the central Swiss Alps were created as a result of isostatic adjustment subsequent to an episode of pre-Maestrichtian folding. During the Maestrichtian to middle Eocene time, the Ultrahelvetic region was characterized by a topography typical of block-faulted terranes. Rugged land mass adjoining deep-sea basins resulted in conditions favorable for periodic turbidity-current deposition in those basins.

(2) The Schlieren Flysch, the Leissigen Flysch, and the Sandstone Flysch near Adelboden were deposited in a row of near-shore deep-sea basins just south of a shallow shelf sea where carbonates of the Helvetic facies accumulated. These Flysch formations derived their sediments from the south or southwest. They were subsequently overthrust as the nappes of the Internal Prealps.

(3) The Wäggital Flysch and the Gurnigel Flysch were deposited in a row of deep-sea basins farther offshore. These were separated from the near-shore basins by an island chain which furnished sediments to both units. Subsequently, the Gurnigel Flysch was overthrust as a nappe of the External Prealps.

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Introduction

The Ultrahelvetic Flysch sandstones of the Swiss Prealps occur as isolated outcrops in a belt that extends from Liechtenstein to Lake Geneva (Fig. 1). In most places these rocks, which range in age from Cenomanian to Eocene (Table 1), were detached from underlying strata during alpine deformation. The very difficult problem of finding the original sites of these "rootless" sediments has been attacked for many years by alpine geologists by means of painstaking accumulation of data and brilliant syntheses. Figuratively the alpine nappes were unrolled and clues uncovered for paleogeographic reconstruction. Such paleogeographic interpretations in the past, however, have been based chiefly on stratigraphic and tectonic analyses. Some questions still need answering and some disputes need settling because of the imperfection of the geologic record. New data from the study of sedimentary structures provide additional information. In this paper some previous syntheses are examined and alternative interpretations are discussed. Crowell (1955) made measurements of the directional sedimentary structures in the Ultrahelvetic Gurnigel Flysch of western Switzerland; the writer hoped in this same way to add paleocurrent data and to add to knowledge of the Late Cretaceous-early Tertiary paleogeography of the Swiss Alps.

In the autumn of 1957 and the spring of 1958, the writer made a field survey of depositional structures in the Flysch of the Swiss Alps. It was found that Ultrahelvetic Flysch formations in other parts of Switzerland have many sedimentary structures in common with the Gurnigel Flysch. Many local names have been given to these various occurrences because of uncertainties in correlation (Fig. 1). In Liechtenstein the Ultrahelvetic zone is probably represented by the Vorarlberg Flysch (Allemann and Blaser, 1950; Richter, 1956), which is considered a correlative of the Ultrahelvetic Flysch of the Austrian Alps (Kraus, 1932; 1942; Tercier, 1936). Traced discontinuously westward the Ultrahelvetic Sandstone Flysch includes the Fäneren Flysch in Appenzell (Eugster and others, 1943), the Wildhaus Sandstone Flysch of St. Gallen (Forrer, 1946), the Wäggital Flysch in Schwyz (Kraus, 1942; Leupold, 1942), the Sardona Flysch in Glarus (Leupold, 1937; 1942), the Schlieren Flysch of central Switzerland (Schaub, 1951), the Gurnigel Flysch of the External Prealps (Tercier, 1928a), and the various Schlieren-Gurnigel-type Sandstone Flysch of the External Prealps and Internal Prealps (Gagnebin, 1924; Huber, 1933). Similar Ultrahelvetic Flysch sandstones are present in the French Alps south of Lake Geneva (Lillie, 1937; 1939). Sedimentary structures indicating the direction of the currents responsible for them are common in the Wäggital Flysch, the Schlieren Flysch, the Gurnigel Flysch, and in the Schlieren-Gurnigel-type Flysch formations of the Internal Prealps and External Prealps.

Interpretation of paleocurrent structures observed in the field confirmed earlier suggestions that the early Tertiary Ultrahelvetic realm of the Swiss Alps was characterized by a topography typical of block-faulted terranes (Tercier, 1939; Crowell, 1955). The Ultrahelvetic basins might have been created as a result of isostatic adjustment subsequent to episodes

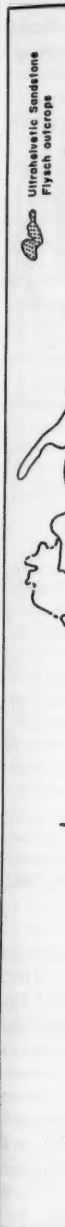

FIGURE 1.—DISTRIBUTION OF THE ULTRAHELVETIC FLYSCH OF THE SWISS PREALPS

A—Sandstone Flysch near Adelboden; C—Cucloz Flysch; F—Fäneren Flysch; G—Gurnigel Flysch; L—Leissigen Flysch; P—Prättigau Flysch (Pennine Flysch); S—Schlieren Flysch; Sa—Sardona Flysch; W—Wäggital Flysch; Wd—Wildhaus Flysch; AR—Aiguilles Rouges massif; MB—Mont Blanc massif; Tav—Tavetscher massif

of local folding of the Pennine Alps during Late Cretaceous. Sedimentary structures of the Flysch sandstones indicate that these sandstones resulted from intermittent turbidity-current deposition in deep-sea basins (Crowell, 1955). Paleocurrent data lend support to previous arguments that the root zone of the Ultrahelvetic nappes might be placed south of the Gotthard massif and that the source materials of the Flysch sandstones were derived from an ancient massif now buried under the Pennine front (Leupold, 1942; Crowell, 1955). Paleogeographic reconstructions also suggest that the Gurnigel Flysch of the External Prealps belongs to a higher tectonic element than the Sandstone Flysch of the Internal Prealps.

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Table 1.—Age Correlation of Ultrahelvetic, Helvetic, and Autochthonous Flysch

		Vorarlberg Flysch[1]	Fäneren Flysch[2]	Wildhaus Flysch[3]	Wäggital Sandstone Flysch[4]	Sardona Flysch[5]	Schlieren Flysch[6]	Sandstone Flysch Internal Prealps[7]	Gurnigel Flysch[8]	Wildflysch s.s.[9]	Subalpine Flysch (Central Switzerland)[10]	Subalpine Flysch (West Switzerland)[11]	North Helvetic and Autochthonous Flysch[12]
Lower Oligocene												↕	↑
Eocene	Upper Priabonian												↓
	Lower Priabonian									↕	↕		
	Lutetian							↑	↑				
	Ypresian						↑	↓	↕				
Paleocene					↑	↑	↕		↕				
Maestrichtian (including Danian)		↑			↕	↕	↕		↕				
Campanian		↕											
Lower Senonian		↕											
Turonian		↓	↕	↑									
Cenomanian				↓									

[1] Allemann, Blaser, and Nänny (1951); [2] Eugster and others (1943); [3] Forrer (1946); [4] Leupold (1942); [5] Leupold (1942); [6] Schaub (1951); [7] Liechti (1931); [8] Guillaume (1957); [9] Leupold (1942); [10] Furrer (1949); [11] Mornod (1949); [12] Oberholzer (1933), Vuagnat (1944)

Pettijohn, Dr. E. Cloos, and Dr. R. C. Murat critically read the manuscript and made many valuable suggestions. However, the author assumes the responsibility for the views and conclusions presented in this paper.

Directional-Current Structures

The types of sedimentary structures that are exhibited by the Prealpine Flysch have been described by Crowell (1955) and are summarized as follows:

On bottom of bed
- Substratal lineation
  - Groove-cast lineation
    - Drag marks
    - Slide marks
  - Load-cast lineation
- Flute cast
- Torose load cast

Inside bed
- Internal lineation
  - Clast lineation
    - Grain lineation
    - Charcoal-fragment lineation
  - Parting lineation
- Current statification (cross-bedding)
- Convolute bedding (slump fold)

On top of bed
- Ripple marks

Flute casts and groove casts are the most common types and are therefore most valuable indicators of the paleocurrent directions of the Ultrahelvetic Flysch. Groove casts are parallel straight ridges on the bottom of a bed, which in most beds is a sandstone. The relief of the ridges, representing the depth of original grooves, ranges from less than 1 mm to several centimeters. These straight ridges or lines are parallel; many extend for meters. Flute cast are sharp conical welts on the undersurfaces of a bed.

"One end of the ridge is rounded or bulbous, and the other flares out to merge gradually with the striated bottom surface of the sandstone layer. Flute casts range in size from a few millimeters in width and about a centimeter in length up to more than 10 cm by 40 cm. The place of maximum relief is usually a short distance from the rounded 'beak,' which rises steeply from the flat undersurface with an overhang....Flute casts are useful in giving direction of flow along the lineation—the beaked or steep end is upstream." (Crowell, 1955, p. 1360).

The writer has pointed out the contrasting origins of the flute casts and groove casts (Hsu, 1959a). Where the turbulent water of turbidity current directly impinged on a muddy sea bottom, subconical depressions were produced which eventually are preserved as flute casts on the undersurface of the overlying sandstone. However, conditions for laminar flow might set in at the bottom of some turbidity current after it has traveled a certain distance; sediment debris dragged along by this laminar motion incised on a muddy sea bottom straight grooves which are eventually preserved as groove casts on the undersurface of the overlying sandstone. Corollaries of this hypothesis are: (1) flute casts are more common in poorly graded or nongraded turbidity-current deposits, and groove casts are more common in graded deposits; (2) flute casts indicate the direction of an earlier stage of turbidity flow than that indicated by groove casts; (3) multiple sets of groove on the undersurface of a sandstone are more common than multiple sets of flute casts; (4) some graded sands should be laminated, and some might contain only a little interstitial clay. All these corollaries are borne out by the writer's observations of the Ultrahelvetic Flysch of the Swiss Alps.

Good exposures of cross-stratifications and convolute bedding were also studied. In general, these indicate the same paleocurrent directions as the flute and groove casts.

Crowell's (1955) method of measuring the angle between the line of strike and the lineation in the plane of bedding is adopted for the present work. The results are tabulated in Table 2. For the sake of clarity the counterclockwise angle, measured on the bottom of the bedding plane, between the dip of the bed and the direction of the source of the transporting current, was recorded. The source direction is assumed to be the opposite of the direction of the current flow. The beaked end of flute casts points upcurrent. The dip of the cross-beds and the direction of overturning of the convolute folds point downcurrent. Groove casts give merely the direction of the current line but not the sense of current flow along this line. Directional observations measured in the field have been rotated around the present strike in order to bring them approximately into the position they must have had when formed. The error so introduced is probably small, as suggested by the fact that the paleocurrent structures in Flysch rocks dipping to various directions show more or less parallel current orientation after

TABLE 2.—MEASUREMENTS OF PALEOCURRENT STRUCTURES, ULTRAHELVETIC FLYSCH, SWISS ALPS

See Figures 2, 5, 10, 11, 14 for locations.

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
Wägital Flysch	1	A	20° S.5° E.	190°	N.5° E.		Groove cast
	2	A	25° S.	185°	N.5° E.		Groove cast
		B		175°	N. 5° W.		Groove cast
		C		165°	N.15° W.		Groove cast
	3	A		140°		N.40° W.	Current stratification
	4	A	22° S. 5° W.	195°		N.20° E.	Flute cast
	5	A		175°		N.	Convolute bedding
	6	A		150°		N.25° W.	Current stratification
		B		185°		N.10° E.	Flute cast
	7	A		185°		N.10° E.	Flute cast
		B		188°	N.13° E.		Groove cast
	8	A	20° S.5° E.	115°		N.70° W.	Flute cast and groove cast
		B		180°		N.5° W.	Flute cast and groove cast
		C		195°		N.10° E.	Flute cast and groove cast
		D		180°		N.5° W.	Convolute bedding
		E		200°		N. 15° E.	Flute cast
		F		185°		N.	Flute cast
	9	A		180°		N. 5° W.	Flute cast
		B		205°	N. 20° E.		Groove cast
	10	A	25° S. 25° E.	210°		N. 5° E.	Flute cast
		B		210°		N. 5° E.	Convolute bedding
	11	A	25° S. 30° W.	145°	N. 5° W.		Groove cast
		B		160°	N. 10° E.		Groove cast
		C		165°		N. 15° E.	Flute cast
	12	A		135°		N. 15° W.	Flute cast
		B		135°		N. 15° W.	Flute cast
		C		145°		N. 5° W.	Flute cast
		D		140°		N. 10° W.	Flute cast and groove cast
	13	A	30° N. 65° W.	70°		N. 5° E.	Flute cast
	14	A		65°		N.	Current stratification
	15	A	20° S. 15° W.	180°		N. 15° E.	Flute cast
		B		190°		N. 25° E.	Flute cast
	16	A	25° S. 25° W.	170°		N. 15° E.	Flute cast
	17	A		155°		N.	Flute cast
		B		165°		N. 10° E.	Convolute bedding
		C		115°		N. 40° W.	Current stratification
		D		155°		N.	Flute cast
		E		160°		N. 5° E.	Current stratification
Schlieren Flysch	1	A	40° S. 60° W.	95°		N. 25° W.	Flute cast
		B		−15°		S. 45° W.	Flute cast
		C	35° S. 35° W.	85°	N. 60° W.		Groove cast
		D	35° S. 45° W.	90°	N. 45° W.		Groove cast
		E		95°	N. 40° W.		Groove cast
		F 1		90°	N. 45° W.		Groove cast
		F 2		65°	N. 70° W.		Groove cast

TABLE 2.—*Continued*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
	2	A	30° S. 55° W.	0°		S. 55° W.	Giant flute cast
		B		−5°		S. 50° W.	Flute cost
		C 1		0°		S. 55° W.	Giant flute cast
		C 2		5°	S. 60° W.		Groove cast
		D		40°	N. 85° W.		Groove cast
		E		0°	S. 55° W.		Groove cast
		F		0°	S. 55° W.		Groove cast
		G		50°		N. 75° W.	Flute cast and groove cast
		H		5°	S. 60° W.		Groove cast
		I		10°	S. 65° W.		Giant groove cast
	3	A		35°	S. 90° W.		Giant groove cast
		B		10°	S. 65° W.		Giant groove cast
		C		10°		S. 65° W.	Flute cast
	4	A		5°	S. 60° W.		Groove cast
		B		10°		S. 65° W.	Flute cast
		C 1		5°	S. 60° W.		Scoured channel
		C 2		15°	S. 70° W.		Scoured channel
	5	A	28° S. 45° W.	25°	S. 70° W.		Giant groove cast
		B		25°	S. 70° W.		Giant groove cast
		C		25°		S. 70°W.	Giant flute cast
		D		30°		S. 75° W.	Flute cast and groove cast
		E		20°		S. 65° W.	Flute cast
		F		−5°	S. 40° W.		Groove cast
Schlieren Flysch	6	A	30° S. 35° W.	10°		S. 45° W.	Flute cast
		B		25°	S. 60° W.		Groove cast
		C		45°	S. 80° W.		Groove cast
		D 1		60°	N. 85° W.		Groove cast No. 1
		D 2		50°	S. 85° W.		Groove cast No. 2
		E		65°		N. 80° W.	Flute cast
		F		65°	N. 80° W.		Groove cast
		G		50°	S. 85° W.		Groove cast
		H		55°	S. 90° W.		Groove cast
		I		50°	S. 85° W.		Groove cast
	7	A	30° S. 40° W.	50°	S. 90° W.		Groove cast
		B 1		45°	S. 85° W.		Groove cast No. 1
		B 2		60°	N. 80° W.		Groove cast No. 2
		B 3		60°	N. 80° W.		Grain lineation
		C		35°		S. 75° W.	Convolute bedding
		D		55°	N. 85° W.		Groove cast
		E		50°	S. 90° W.		Groove cast
		F 1		60°	N. 80° W.		Groove cast
		F 2		70°	N. 70° W.		Groove cast
		F 3		40°	S. 80° W.		Groove cast
		G		70°		N. 70° W.	Flute cast
	8	A 1		45°	S. 85° W.		Groove cast No. 1
		A 2		35°	S. 75° W.		Groove cast No. 2
		B		30°		S. 70° W.	Flute cast
		C 1		15°	S. 55° W.		Groove cast No. 1
		C 2		−5°	S. 35° W.		Groove cast No. 2
		C 3		−15°	S. 25° W.		Groove cast No. 3

TABLE 2.—*Continued*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
Schlieren Flysch	9	A 1		20°		S. 60° W.	Flute cast
		A 2		5°	S. 45° W.		Groove cast
		B		50°	S. 90° W.		Giant groove cast
		C		−10°		S. 30° W.	Flute cast
	10	A		45°	S. 85° W.		Groove cast
		B		35°	S. 75° W.		Giant groove cast
		C 1		70°	N. 70° W.		Groove cast No. 1
		C 2		55°	N. 85° W.		Groove cast No. 2
		D		60°	N. 80° W.		Groove cast
		E		65°	N. 75° W.		Giant groove cast
	11	A 1		55°	N. 85° W.		Groove cast No. 1
		A 2		35°	S. 75° W.		Groove cast No. 2
		B		35°	S. 75° W.		Groove cast
		C		55°	N. 85° W.		Giant groove cast
		D		35°	S. 75° W.		Giant groove cast
	12	A 1		55°	N. 85° W.		Groove cast No. 1
		A 2		65°	N. 75° W.		Groove cast No. 2
		B		35°	S. 75° W.		Groove cast
	13	A		70°	N. 70° W.		Groove cast
	14	A 1	35° S. 30° W.	40°	S. 70° W.		Groove cast
		A 2		60°	S. 90° W.		Groove cast
	15	A		40°	S. 70° W.		Groove cast
		B		0°		S. 30° W.	Flute cast and groove cast
		C		65°	N. 85° W.		Groove cast
		D		15°		S. 45° W.	Giant flute cast
		E		50°	S. 80° W.		Groove cast
	16	A 1		25°	S. 55° W.		Groove cast No. 1
		A 2		40°	S. 70° W.		Groove cast No. 2
		A 3		50°	S. 80° W.		Groove cast No. 3
		B 1		35°	S. 65° W.		Groove cast No. 1
		B 2		45°	S. 75° W.		Groove cast No. 2
		C		55°	S. 85° W.		Groove cast
	17	A 1		60°	S. 90° W.		Groove cast No. 1
		A 2		20°	S. 50° W.		Giant groove cast No. 2
		B		55°	S. 85° W.		Groove cast
		C 1		70°	N. 80° W.		Groove cast No. 1
		C 2		50°	S. 80° W.		Groove cast No. 2
	18	A 1		35°	S. 65° W.		Groove cast
		A 2		25°		S. 55° W.	Flute cast
	19	A		30°	S. 70° W.		Groove cast
	20	A	60° S. 60° W.	15°		S. 75° W.	Flute cast
		B		15°		S. 75° W.	Current stratification
		C		55°	N. 65° W.		Groove cast
		D		10°		S. 70° W.	Flute cast
	21	A 1		60°	N. 60° W.		Groove cast No. 1
		A 2		45°	N. 75° W.		Groove cast No. 2
		B		65°	N. 55° W.		Groove cast
		C		5°	S. 65° W.		Groove cast
	22	A	55° S. 40° W.	15°		S. 55° W.	Flute cast

TABLE 2.—*Continued*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
	23	A 1	60° S.	60°		S. 60° W.	Flute cast No. 1
		A 2		130°	N. 50° W.		Groove cast
		B		60°	S. 60° W.		Groove cast
				120°	N. 60° W.		Groove cast
	24	A 1		75°	S. 75° W.		Groove cast No. 1
		A 2		100°	N. 80° W.		Groove cast No. 2
		A 3		125°	N. 65° W.		Groove cast No. 3
		B		85°	S. 85° W.		Groove cast
		C 1		90°	S. 90° W.		Groove cast No. 1
		C 2		105°	N. 75° W.		Groove cast No. 2
		C 3		120°	N. 60° W.		Groove cast No. 3
		C 4		85°		S. 85° W.	Flute cast
		D		85°	S. 85° W.		Groove cast
	25	A	60° S. 25° E.	110°		S. 85° W.	Flute cast
		B		105°		S. 80° W.	Flute cast
		C	60° S. 75° E. (overturned)	−10°		S. 85° W.	Flute cast
		D	80° S. 25° E. (overturned)	90°		S. 65° W.	Flute cast
	26	A	50° S. (overturned)	−60°		S. 60° W.	Flute cast
		B	45° S. 25° E. (overturned)	90°		S. 65° W.	Flute cast
Schlieren Flysch	27	A	30° N. 10° E.	−110°		S. 80° W.	Flute cast
	28	A	30° S.	110°	N. 70° W.		Groove cast
		B		75°	S. 75° W.		Groove cast
		C		105°	N. 75° W.		Groove cast
		D		70°	S. 70° W.		Groove cast
		E		90°	S. 90° W.		Groove cast
		F		100°	N. 80° W.		Groove cast
		G		135°	N. 45° W.		Groove cast
		H		130°	N. 50° W.		Groove cast
		I 1		105°	N. 75° W.		Groove cast No. 1
		I 2		115°	N. 65° W.		Groove cast No. 2
		J		120°	N. 60° W.		Groove cast
		K		135°	N. 45° W.		Groove cast
		L		95°		N. 85° W.	Flute cast
	29	A	50° N. 15° W.	75°		S. 90° W.	Flute cast
		B		120°	S. 45° W.		Groove cast
		C		115°		S. 50° W.	Flute cast
		D		120°	S. 45° W.		Groove cast
	30	A	50° N. 10° W.			S. 80° W.	Flute cast
		B 1			S. 75° W.		Groove cast No. 1
		B 2			S. 60° W.		Groove cast No. 2
		C			S. 90° W.		Groove cast
		D			S. 70° W.		Groove cast
	31	A	40° S. 45° E.	140°	N. 85° W.		Groove cast (?)
		B		145°	N. 80° W.		Groove cast
		C		145°	N. 80° W.		Groove cast

TABLE 2.—*Continued*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
Schlieren Flysch	32	A		125°	S. 80° W.		Groove cast
		B		135°		S. 90° W.	Flute cast
		C		115°	S. 70° W.		Groove cast
		D		135°	S. 90° W.		Groove cast
		E		120°	S. 75° W.		Groove cast
		F		120°		S. 75° W.	Flute cast (?)
		G		135°		S. 90° W.	Flute cast
		H		135°	S. 90° W.		Groove cast
		I		110°	S. 65° W.		Groove cast
		J		115°	S. 70° W.		Groove cast
		K		135°		S. 90° W.	Flute cast
		L		185°	N. 40° W.		Groove cast
		M		130°	S. 85° W.		Groove cast
		N		130°	N. 85° W.		Groove cast
	33	A 1	40° S. 20° E.	135°	N. 65° W.		Groove cast No. 1
		A 2		160°	N. 40° W.		Groove cast No. 2
	34	A	40° S. 40° E.	90°		S. 50° W.	Flute cast
		B		90°		S. 50° W.	Flute cast
		C		90°	S. 50° W.		Groove cast
		D		90°		S. 50° W.	Flute cast
		E		95°		S. 55° W.	Flute cast
		F		110°		S. 70° W.	Flute cast
		G		135°	N. 85° W.		Groove cast
		H		115°	S. 75° W.		Groove cast
		I		130°	S. 90° W.		Groove cast
		J		90°		S. 50° W.	Flute cast
		K		145°	N. 75° W.		Groove cast
		L	25° S. 75° E.	180°	N. 75° W.		Groove cast
		M 1		150°	S. 75° W.		Groove cast
		M 2		165°		S. 90° W.	Flute cast
		N	20° N. 65° E.	255°		N. 80° W.	Flute cast
		O		260°		N. 75° W.	Flute cast
	35	A	30° N. 30° E.	−110°		N. 80° W.	Flute cast
		B		−95°		N. 65° W.	Flute cast
		C		−85°		N. 55° W.	Flute cast
		D		−130°	S. 80° W.		Groove cast
		E		−75°	N. 45° W.		Groove cast
		F		−120°	S. 90° W.		Groove cast
		G		−85°	N. 55° W.		Groove cast
		H		−100°	N. 70° W.		Groove cast
		I		−100°	N. 70° W.		Groove cast
		J		−75°		N. 45° W.	Flute cast and groove cast
		K		−110°	N. 80° W.		Groove cast
	36	A		−100°	N. 70° W.		Groove cast
		B		−80°	N. 50° W.		Groove cast
		C		−125°		S. 85° W.	Giant flute cast
		D		−70°		N. 40° W.	Flute cast
	37	A	30° S. 65° W.	25°		S. 90° W.	Flute cast (?)
		B	70° S. 50° W.	20°		S. 70° W.	Flute cast (?)
	38	A	30° S. 10° W.	45°	S. 55° W.		Groove cast
	39	A	50° N.	180°		S.	Flute cast (?)

TABLE 2.—*Continued*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
Leisigen Flysch	1	A	80° N. 70° W.	−30°	S. 80° W.		Groove cast
	2	A	50° N. 55° W.	−135°		S. 10° E.	Current stratification
		B		−150°		S. 25° E.	Convolute bedding
		C		−130°		S. 5° E.	Current stratification
		D	30° N. 80° W.	−85°		S. 15° W.	Flute cast
		E	40° N. 50° W.	−120°		S. 10° W.	Current stratification
		F		−115°		S. 15° W.	Current stratification
		G	55° N. 75° W.	−100°		S. 5° W.	Flute cast
	2	A 1		−60°	S. 45° W.		Groove cast (?) (striations)
		A 2		−105°		S.	Current stratification
		B	50°–80° N. 45° W.	−100°	S. 35° W.		Groove cast (?) (striation)
		C		−90°	S. 45° W.		Groove cast (?) (striations)
		D		−135°	S. 10° W.		Groove cast
		E		−100°		S. 35° W.	Flute cast
		F		−85°		S. 50° W.	Flute cast
		G		−90°		S. 45° W.	Current stratification
		H		−85°		S. 50° W.	Flute cast
		I		−90°		S. 45° W.	Flute cast
		J		−110°		S. 25° W.	Flute cast
		K		−60°		S. 75° W.	Flute cast
		L		−100°		S. 35° W.	Flute cast
		M	85° N. 65° W.	−20°		N. 85° W.	Flute cast
	3	A	70° N. 65° W.	−30°		S. 85° W.	Flute cast
		B		−80°		S. 35° W.	Flute cast
	4	A	85° N. 30° W.	−95°		S. 55° W.	Flute cast
		B		−105°		S. 45° W.	Flute cast
		C		−120°		S. 30° W.	Flute cast
		D		−170°	S. 40° W.		Groove cast
		E		−130°		S. 20° W.	Flute cast
		F	80° N. 40° W.	−100°		S. 40° W.	Giant flute cast
		G		−105°		S. 35° W.	Flute cast
	5	A	50° S. 30° E.	90°		S. 60° W.	Flute cast
		B		75°		S. 45° W.	Flute cast
	6	A	70° S. 45° E.	85°		S. 40° W.	Flute cast
		B		85°		S. 40° W.	Flute cast
Flysch Internal Prealps	1	A	65° N. 80° W. (overturned)	−90°		S. 10° W.	Flute cast
	2	A	50° N. 75° E.	150°	S. 45° W.		Groove cast (?)
		B		110°		S. 5° W.	Flute cast (?)
	3	A1	30° S. 60° E.	105°		S. 45° W.	Flute cast No. 1
		A2		130°		S. 70° W.	Flute cast No. 2
		B		125°		S. 65° W.	Current stratification
	4	A 1	15° N. 65° W.	−90°		S. 25° W.	Giant flute cast
		A 2		−110°		S. 5° W.	Flute cast

TABLE 2.—*Concluded*

Formation	Location	Observation number	Dip of bedding	Angle between lineation and dip direction	Bearing of lineation	Bearing to source of current	Type of directional-current structure
Gurnigel Flysch	1	A	55° S. 30° E.	105°		N. 45° E.	Flute cast
		B		110°	N. 40° E.		Charcoal fragment lineation
		C		105°		N. 45° E.	Flute cast
		D		105°		N. 45° E.	Flute cast
		E		115°	N. 35° E.		Groove cast
		F		120°	N. 30° E.		Groove cast
	2	A		115°		N. 35° E.	Current stratification
		B 1		130°		N. 20° E.	Current stratification
		B 2		140°		N. 10° E.	Current stratification
		B 3		140°		N. 10° E.	Current stratification
		C		110°		N. 40° E.	Flute cast

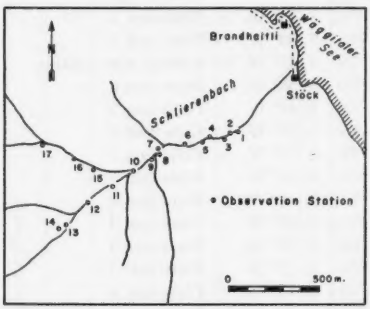

FIGURE 2.—SKETCH MAP SHOWING LOCATION OF OBSERVATION STATIONS IN THE WÄGGITAL FLYSCH

The Lachen sheet (Blatt 236), *Landeskarte der Schweiz*, was used as the base map.

being rotated back into horizontality.[1] After rotation the paleocurrent data were plotted on the current-rose diagrams, which were first devised by Crowell (1955) to aid in paleogeographic interpretations. Lineations where so plotted extend through the center of the plot and into two quadrants. Flute casts, crossbedding, and direction of overturning of convolute bedding are confined to only one quadrant.

[1] Locally, the Prealpine Flysch might have been rotated horizontally during the alpine deformation. (*See* Heim, 1932; Buxtorf, 1936; Renz, 1937.) The Schlieren Flysch of the Gross Schliere strikes southeast, almost at right angles to the regional east-northeast trend. A correction for such a horizontal rotation, however, would strengthen rather than weaken the main conclusion of this paper. (*See* footnote 5).

WÄGGITAL SANDSTONE FLYSCH

Name, Age, and Distribution

The term Wäggital Flysch has been used by Leupold (1942) to describe the Wildflysch and the Sandstone Flysch on the backs of the main Helvetic nappes of the Glarus Alps (Fig. 1). The Flysch extends from the vicinity of the Wäggitaler See westward to Sattel-Schwyz Road. However, Kraus (1932) noted that the distribution of the sandstone facies of the Wäggital Flysch is restricted to the west side of Wäggitaler See in the drainage basin of the Schlierenbach (Fig. 2). On microfaunal evidence, Leupold assigned a Maestrichtian age for the Wäggital Sandstone Flysch exposed at the Schlierenbach on the west side of the Wäggitaler See. He further suggested that the upper part of the Flysch might be Paleocene.

The structure of the Glarus Alps is shown in Figure 3. The sequence of the nappes is listed as follows in decending order (Buxtorf, 1957):

V Klippes
  Mesozoic sedimentary rocks
IV Ultrahelvetic or South Helvetic zone
  Sandstone Flysch (Wäggital, Sardona) Maestrichtian to Paleocene

FIGURE 3.—TECTONIC CROSS SECTION THROUGH THE GLARUS ALPS (ABOVE) AND THROUGH THE CENTRAL SWISS ALPS (BELOW)
After Buxtorf (1957)

Wildflysch, Priabonian?
Einsiedeln and Blattengrat Suchuppen (enveloped South Helvetic zone, Paleocene to Eocene)
III Helvetic nappes
Drusberg-Säntis nappe (Wildhorn nappe)
Axen nappe
Mürtschen nappe
II Parautochthonous units
I Hercynian massifs and their autochthonous mantle

Tectonically, the Wäggital Flysch occupies a position below the Pennine metamorphic rocks and above the South Helvetic Einsiedeln and Blattengrat Schuppen. Kraus (1932) postulated that the Wäggital Flysch consists of two units. A Sandstone Flysch has been thrust over a Wildflysch of a different age and origin. Recent work by Allemann, Blaser, and Nänny (1951) supported this interpretation. During the course of the present work the difference in sedimentary structures between the Sandstone Flysch and the Wildflysch was noted. Flute casts are abundant in the sandstone Flysch facies, and they are practically absent in the Wildflysch facies.

Lithology

Measurements were obtained from the Wäggital Sandstone Flysch exposed at Schlierenbach. The section consists predominantly of calcareous sandstone and shale. Other rock types identified include argillaceous sandstone, calcareous siltstone, and dense impure limestone. The sandstone, in general very fine-grained, exhibits sedimentary structures typical of turbidity-current deposits (Kuenen and Migliorini, 1950; Kuenen and Carozzi, 1953; Crowell, 1955); graded bedding, groove casts, and flute casts are common. The calcareous siltstone layers rarely display graded bedding. Cross-laminations and convolute bedding are common. On the bottom of siltstone beds, fucoids, helminthoids, or flute casts are present, but groove casts and load casts have not been found. Some cryptocrystalline, tough, argillaceous limestone beds are present. Flute casts have been found in some of these light-gray impure limestone beds. One of the limestone beds was deposited in a shallow channel scoured out of the underlying shales. The shale is laminated or schistose. Some laminations show small-scale graded bedding with the lower part more silty. Others are not graded. Substratal lineations and flute casts are absent on the bottom of shale beds.

Direction of Transport

About 40 measurements of directional current structures were made for the Wäggital Flysch. The results are consistent. The currents responsible for the deposition of the Wäggital Sandstone Flysch in Schlierenbach flowed from north to south (Fig. 4).

Paleogeographic Implications

The Wäggital Flysch, detached from its root and now thrust over the main Helvetic nappes, belongs to the group of sediments which have been generally designated Ultrahelvetic Flysch. The tectonic position of the Wäggital Flysch and its correlation with the other Ultrahelvetic Flysch units are still uncertain. Leupold (1942) assigned both the Wäggital and the Sardona Flysch to northern Pennine elements and correlated them with the Prätigau Flysch. He envisioned a north-south depositional strike and postulated a gradual eastward and possibly southeastward deepening of Maestrichtian-Paleocene sea from the Helvetic shelf to the Pennine deep. Kraus (1932) correlated the Wäggital Flysch with the Upper Cretaceous Fäneren Flysch and Vorarlberg Flysch and assigned the Sandstone Flysch facies to his "Sigiswanger Decke" and the Wildflysch facies to his "Wildflysch Decke." Thus, he implied an east-west depositional strike. Nänny (1946) asserted that the Wäggital and Sardona Flysch could not be correlated to the Prätigau Flysch. The former, being Ultrahelvetic, was separated from the Pennine Prätigau Flysch by a swell or a barrier. More recently, Kraus and Nänny's ideas were confirmed by Allemann, Blaser, and Nänny (1951). According to their view, the Wäggital Flysch was laid down in an Ultrahelvetic trough, which extended from Sihltal in Switzerland to Allgäu in Austria. This trough was separated from the southern Helvetic region by a swell. Wäggital Flysch, Wildhaus Flysch, Fäneren Flysch, and Vorarlberg Flysch, though somewhat different in age, were all deposited in the same basin and subsequently formed the same tectonic unit. This Ultrahelvetic basin was separated from the northern Pennine Prätigau basin by one or more swells or barriers.

The direct implication of the study of sedimentary structures is that a land area, possibly an island, must have existed somewhere to the north of the Wäggital Flysch basin to supply sediments to the southerly flowing currents. Allemann, Blaser, and Nänny's contention that

the Wäggital Flysch basin was limited on the north by a swell is affirmed. However, whether the Wäggital basin extended eastward to

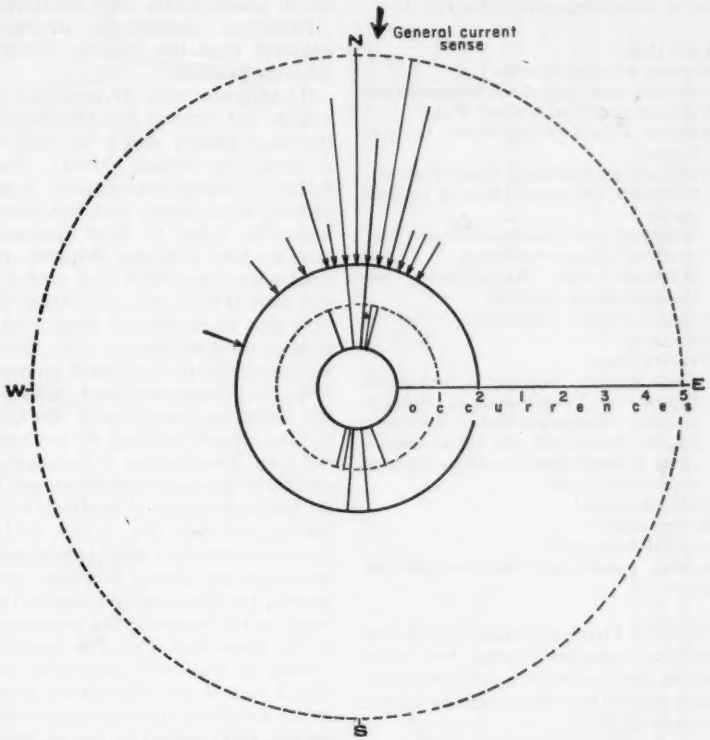

FIGURE 4.—CURRENT-ROSE DIAGRAM, WÄGGITAL FLYSCH, 38 MEASUREMENTS

Vorarlberg or Allgäu and westward to Lucern or Bern is questionable. The Wäggital Sandstone Flysch is Maestrichtian (and possibly also Paleocene) in age (Leupold, 1942), thus younger than the Cenomanian-Turonian Ultrahelvetic Flysch of the Wildhaus and Fäneren, and there has always been the problem of a wide gap in exposures between the Wäggital Flysch and its supposed equivalent, *i.e.*, the Maestrichtian Fanola series of the Vorarl Flysch. (*See* Table 1.) Available data are not sufficient to prove a continuous extension of a Late Cretaceous Ultrahelvetic trough from Vorarlberg to Wäggital.

Although Kraus and others related the Wäggital Flysch to the Vorarlberg Flysch, the similarity of the Wäggital Flysch to the Schlieren Flysch and other Ultrahelvetic Flysch formations of western Switzerland was also noted (*e.g.*, Tercier, 1936; Leupold, 1942). Buxtorf (1957) mentioned that new research now being done by Hans Schaub might give a correlation of the Wäggital Flysch with the Schlieren Flysch. Tectonic correlation of these two Flysch formations is not supported by paleocurrent data (see below).

SCHLIEREN FLYSCH

Name and Age

The Flysch sediments of central Switzerland, *i.e.*, the Schlieren Flysch and the various Wildflysch with exotic blocks of Habkern granite, Leimernschichten, and/or other rocks, have been the subject of many researchers (Kaufmann, 1886; Schider, 1913; Lugeon, 1916; Buxtorf, 1918; Tercier, 1928b; Liechti, 1932; Kraus, 1932; Leupold, 1933; 1942; Vonder-

schmitt and Schaub, 1943; Schaub, 1951; Gigon, 1952). The tectonic framework of central Switzerland, now generally accepted, is outlined below. The nappe and Schuppen elements are listed in descending order. (*See* Fig. 3.)

V Klippen Decke
  Mesozoic sedimentary rocks
IV Ultrahelvetic zone (nappes and Schuppen zone)
  Wildflysch under the Klippen Decke
  Schlieren Flysch, Maestrichtian to lower Eocene
  Wildflysch, *s.s.* Priabonian (upper Eocene)
    Wildflysch with exotic blocks of Habkern granite
    Wildflysch with Leimernschichten "intercalations" (Leimer Schuppen)
    Wildflysch with Wangschichten and Complanatakalk *Schüflinge*
  Subalpine Flysch[2], Priabonian
III Helvetic zone
  Wildhorn nappe
    Early Tertiary (including a conformable sequence of Complanatakalk, Hohgantschiefer, Hohgantsandstein, Schimbergschiefer, Stadschiefer, the last of which is dated as Priabonian) overlying Mesozoic sedimentary rocks
  Diablerets nappe
  Morcles nappe
II Parautochthonous units
I Hercynian massifs and their autochthonous mantles

The Schlieren Flysch has been detached from its underlying sediments during the Alpine deformation and represents a series of continuous sedimentation from Maestrichtian to lower Eocene (Schaub, 1951).

Lithology and Description of Section

The sandstone beds of the Schlieren Flysch all exhibit sedimentary structures characteristic of turbidity-current deposition. Graded bedding is almost invariably present. Substratal lineations such as groove casts and flute casts are both very common. Convolute bedding and cross-stratification are present in finer sediments. They are most common in the calcareous siltstone of the basal Schlieren. The Schlieren sandstones contain pink Habkern-type granite debris (Schider, 1913), although it is somewhat less abundant than in the Gurnigel sandstones. Most of the Schlieren sandstones are coarser and less calcareous than the Wäggital sandstones, and they contain more groove casts and fewer flute casts. The better sorted Schlieren sandstones contain little interstitial clay; they are in places friable after weathering.

Directional sedimentary structures were measured from the Schlieren Flysch at the following localities:

(1) Outcrops west of Alpnachtal in Gross Schliere and vicinity: The Schlieren Flysch of the Gross Schliere section has been described in detail by Schaub (1951). The basal Schlieren consists predominantly of calcareous siltstone which shows abundant slump structures (Fig. 5, loc. 1). Flute casts and groove casts are rare. The lower Schlieren consists of conglomerate, sandstone with graded bedding, and shale (Fig. 5, locs. 2–5). Giant flute casts more than 10 cm wide are common on the bottoms of the conglomerate beds. Groove casts are common on the bottoms of sandstone beds. A beautifully exposed scoured channel is present just behind the waterfall at P. 980[3] (Fig. 5, loc. 4). The channel is about 60 feet wide and 3 feet deep. The direction of the channel wall is parallel to the groove-cast lineations. The middle Schlieren consists of sandstone with graded bedding and shale (Fig. 5, locs. 6–13). Where a sandstone overlies a shale, groove casts and/or flute casts are almost invariably present. In general, the flute casts are smaller than those found on the bottom of the conglomerate beds of the lower Schlieren. The upper Schlieren consists of shale with interbedded sandstone (Fig. 5, locs. 14–19). Groove casts are abundant on the bottom of those sandstone beds, which are also characterized by graded bedding.

Additional data on current directions in this area were obtained from the outcrops of lower Schlieren Flysch at Guber Quarry (Fig. 5, locs. 20, 21) and from the middle and upper Schlieren outcrops on the Alpnach-Horweli Road (Fig. 5, locs. 22–24).

(2) Entlebuch-Sarnen Road: The basal Schlieren Flysch exposed on the side of the Rotbach south of Entlebuch near Rossweid consists of calcareous siltstone alternating with breccia and shale (Fig. 5, locs. 25–27; Schaub, 1951, p. 47). Flute casts were found on the bottom of some siltstone layers. These rocks

[2] The tectonic position of the Subalpine Flysch is still somewhat uncertain. Mornod (1949) considered the Subalpine Flysch of the western Swiss Prealps a sub-Helvetic or a parautochthonous element.

[3] P. 980 and similar expressions are used in this paper to define localities precisely. The number after P. indicates the elevation of the point in meters as read from the base map. P. 980, for example, is located on the base map—in this case the Stans sheet of the Swiss National Topographic Map (*Stans, Blatt 245, Landeskarte der Schweiz*)—where the Gross Schliere creek crosses the 980-m contour.

FIGURE 5.—SKETCH MAP SHOWING LOCATION OF OBSERVATION STATIONS IN THE SCHLIEREN FLYSCH
The Escholzmatt sheet (Blatt 244) and the Stans Sheet (Blatt 245), *Landeskarte der Schweiz*, were used as the base map.

have been very severely deformed, but the current structures show consistent orientation after being rotated back into a horizontal plane.

An excellent exposure of the Schlieren Flysch is present just east of Glaubenberg at an abandoned quarry on the roadside at P. 1449 (Fig. 5, loc. 28). The rocks are interbedded sandstone and shale. Groove-cast lineations are very common on the bottom of the graded sandstones.

Another good exposure of the Schlieren Flysch is present farther east on the highway, at a roadcut just beside the intersection of the highway with the path to Schwendi Kaltbad at P. 1432 (Fig. 5, locs. 29, 30). Flute casts are common there.

(3) Area east of Flühli: The Schlieren sandstone is well exposed on the side of the Rotbach near the bridge on the Flühli-Stäldeli road at P. 1256 (Fig. 5, locs. 31–33). On the bottom of the Flysch sandstone beds are groove-cast lineations and flute casts, which have been much obliterated by weathering. North of Stäldeli, on the bank of the Seewenbach below a waterfall, are some excellent exposures of the Schlieren Flysch (Fig. 5, locs. 34–36; Schaub, 1951, p. 57). The lower part of the section is basal Schlieren and consists of beds of calcareous sandstone with flute casts alternating with shale. The upper part of the exposed section (just below the waterfall) belongs to the lower Schlieren and consists predominantly of sandstone with thin shale intercalations. Groove casts and giant flute casts, like those found in lower Schlieren Flysch at Gross Schliere, are present on the bottom of very coarse sandstones.

(4) Other areas: South of Lake Lucern and east of Alpnachtal, several isolated outcrops of Schlieren Flysch occur (Geiger, 1956). They are present below the Klippen Decke and above the Priabonian Wildflysch of the Habkern zone, which, in turn, has been thrust over the main Helvetic nappes. Only the basal Schlieren Flysch of Maestrichtian age was identified.

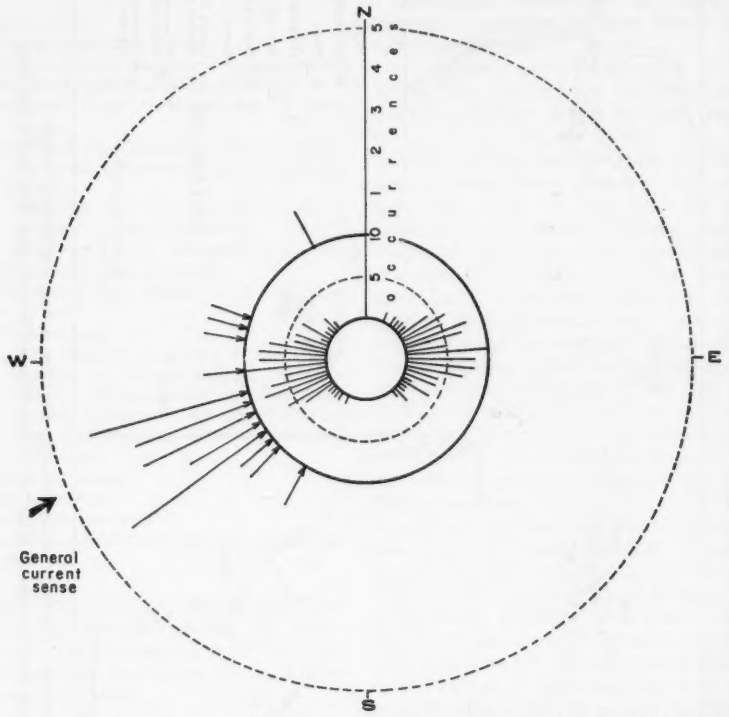

FIGURE 6.—CURRENT-ROSE DIAGRAM, GROSS SCHLIERE SECTION, SCHLIEREN FLYSCH, 117 MEASUREMENTS

These Flysch rocks have been so severely deformed that sedimentary structures have been obliterated in most places. Questionable flute casts were found on the side of the Dallenwil-Wiesenberg road at P. 803 (Fig. 5, loc. 37).

Isolated outcrops of the Schlieren Flysch are present west of the Sarner See (Bentz, 1948; Vonderschmitt and others, 1951). West of Gilswil, on the south bank of the Gross Laui near Abendsitli, groove-cast lineation is present on the bottom of a Schlieren sandstone bed (Fig. 5, loc. 38).

West of Sörenberg, on the cliff face south of the Nünalpstock is well-exposed Schlieren Flysch (Fig. 5, loc. 39; Schaub, 1951, p. 59). Some flute casts were found at the foot of the cliff-face exposure.

Outcrops of Schlieren Flysch on the top of the Schwarzenegg (Fig. 5) south of Sörenberg (Soder, 1949) are somewhat weathered, but groove casts and flute casts were not found.

Direction of Transport

Flute casts are more common on the bottoms of muddy and calcareous sandstone beds which show indistinct or no graded bedding, and giant flute casts are found only on the bottoms of coarse conglomerate layers. Groove casts are more common on the bottoms of sandstones which show distinct graded bedding and have been somewhat sorted. On the whole the currents transporting the sediments of the Schlieren Flysch flowed from south of west to north of east. The average current direction indicated by the flute casts is slightly different from that indicated by the groove casts. For example, the average current direction as indicated by the flute casts in the Gross Schliere section is N.65°E., whereas the average direction as indicated by the groove casts is almost due east (Fig. 6). Flute casts might indicate the direction of the turbidity flows in an earlier stage of transport when the currents were

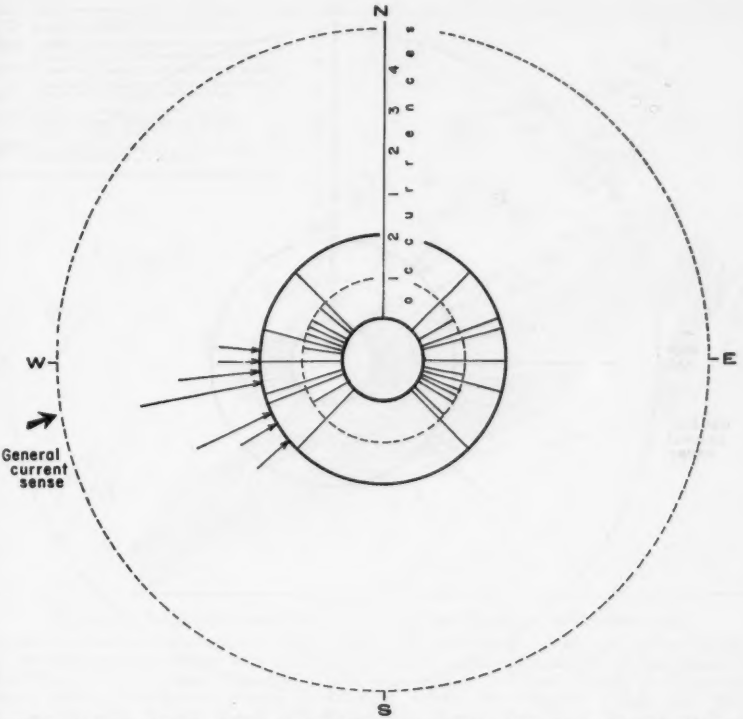

FIGURE 7.—CURRENT-ROSE DIAGRAM, ENTLEBUCH-SARNEN ROAD SECTION, SCHLIEREN FLYSCH, 29 MEASUREMENTS

flowing down the side of a basin (Hsu, 1959a). Sandstones that have groove casts were deposited after these flows had changed their flow directions somewhat on account of submarine topography.

Two-hundred and five measurements of directional sedimentary structures from the Schlieren Flysch were made from 39 localities. The results are tabulated in Table 2 and summarized in Figures 6–8.

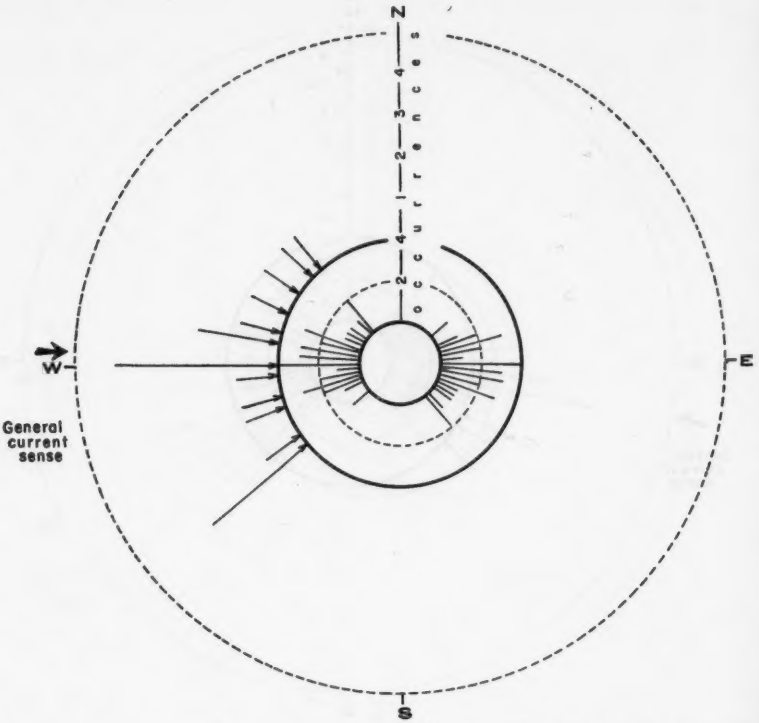

FIGURE 8.—CURRENT-ROSE DIAGRAM, FLÜHLI SECTION, SCHLIEREN FLYSCH, 50 MEASUREMENTS

Paleogeographic Implications

The tectonic position of the rootless Schlieren Flysch has not been ascertained. Because of its position and similarities in age and lithology, the Schlieren Flysch is generally correlated with the Gurnigel Flysch of the External Prealps and with the Gurnigel-Schlieren-type Sandstone Flysch in the Ultrahelvetic zone of the Internal Prealps. These Flysch sandstones, together with some of the Sandstone Flysch of northeastern Switzerland, are thus generally considered Ultrahelvetic elements[4].

The correlation of the Schlieren Flysch with the Wäggital Flysch has not yet been proven by paleontological evidence. Sedimentological data refute a direct connection between these two Flysch formations. The currents depositing the Wäggital Flysch flowed from north to south, whereas the currents depositing the Schlieren

[4] Leupold (1942) considered the Schlieren Flysch a northern Pennine element. He envisioned a swell which separated the trough of the Schlieren Flysch deposition from the southern Helvetic region. However, Leupold's different interpretation is partly merely a difference in terminology. He omitted the term Ultrahelvetic in his scheme, but the Schlieren Flysch was still correlated with other Ultrahelvetic elements (Leupold's northern Pennine) such as the Wäggital Flysch and Sardona Flysch of the Glarus Alps, the Gurnigel Flysch of the External Prealps, and Laubhorn Decke of the Internal Prealps.

Flysch flowed from west to east. They were obviously derived from different source areas and probably deposited in different basins.

The Schlieren Flysch has been correlated with the Gurnigel Flysch on petrographic grounds. The pink Habkern-type granite is a dominant clastic component in both the Schlieren and the Gurnigel Flysch (Tercier, 1928*a*; Schider, 1913). However, such petrographic criteria indicated only that the two Flysch formations had a common source area—a direct connection of the two is not warranted. In fact, such a correlation has never been satisfactory because of age discrepancies (Table 1). Flysch sedimentation in the Gurnigel basin continued as late as upper middle Eocene or upper Lutetian, whereas the youngest Schlieren Flysch is found in lower Eocene or Ypresian (Schaub, 1951). Paleocurrent data indicate that these two Flysch formations were probably deposited in two basins on the opposite sides of a land barrier. The currents transporting the Schlieren Flysch flowed eastward or eastnortheastward from land area characterized by outcropping Habkern granite, whereas the currents transporting the Gurnigel Flysch sediments generally flowed southward from the same source area.[5] This land barrier separating the Schlieren basin from the Gurnigel basin was characterized by a Habkern-type granite basement and might thus be designated as the Habkern massif (Fig. 9).

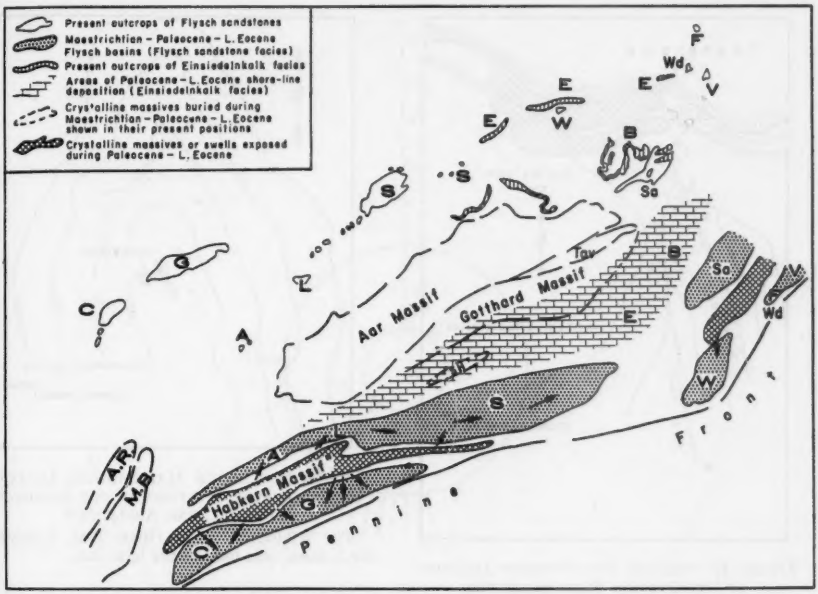
Present outcrops of Flysch sandstones
Maestrichtian – Paleocene – L. Eocene Flysch basins (Flysch sandstone facies)
Present outcrops of Einsiedelnkalk facies
Areas of Paleocene – L. Eocene shore-line deposition (Einsiedelnkalk facies)
Crystalline massives buried during Maestrichtian – Paleocene – L. Eocene shown in their present positions
Crystalline massives or swells exposed during Paleocene – L. Eocene
Aar Massif
Gotthard Massif
Habkern Massif
Penninic Front
Tav
Wd
SH
AR
MB

FIGURE 9.—PALEOGEOGRAPHIC MAP, HELVETIC AND ULTRAHELVETIC SWISS ALPS

Compare the location of the ancient Flysch basins to the position of the outcrops of the Ultrahelvetic Flysch: the displacement resulted from overthrusting during Alpine deformation. A—Sandstone Flysch near Adelboden B—Blattengrat formation (Carbonate facies); C—Cucloz Flysch; E—Einsiedelnkalk (Carbonate facies); F—Fäneren Flysch; G—Gurnigel Flysch; W—Wäggital Flysch; Wd—Wildhaus Flysch; AR—Aiguilles Rouges massif; MB—Mont Blanc Massif; SH—South Helvetic swell; Tav—Tavetscher Massif.

ULTRAHELVETIC FLYSCH OF THE INTERNAL PREALPS

Name, Age, and Distribution

South of Thunersee in the vicinity of Leissigen, Ultrahelvetic Flysch sandstones are

[5] If a clockwise horizontal rotation during the alpine deformation is assumed for the Schlieren Flysch of the Gross Schliere, the corrected directions of transport for the Schlieren Flysch would be more northerly, thus contrasting even more strongly to the southerly direction of transport for the Gurnigel Flysch.

present (Fig. 1; Liechti, 1931). A local name, "Leissigen Flysch," has been given these sediments (Beck, 1928; Liechti, 1931), but the name is not in general use. Similar Ultrahelvetic

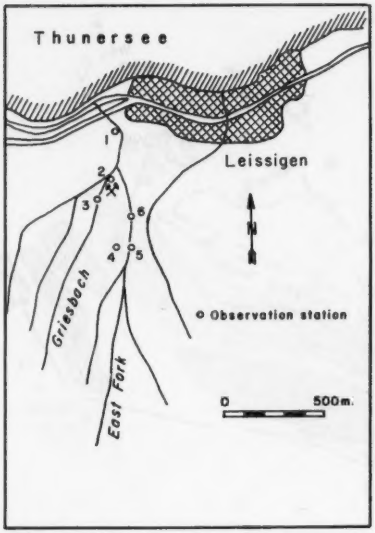

FIGURE 10.—SKETCH MAP SHOWING LOCATION OF OBSERVATION STATIONS IN THE LEISSIGEN FLYSCH

The cross-hatched area indicates the extent of the village of Leissigen. The Interlaken sheet (Blatt 254), *Landeskarte der Schweiz*, was used as the base map.

Flysch sandstones have been found farther southwest in the Sattelzone near Adelboden (Huber, 1933). Both of these Flysch formations have been mapped or referred to as the Schlieren Flysch (Huber, 1933; Leupold, 1942). They are grouped under the Gurnigel-Schlieren-type Ultrahelvetic Flysch in the new 1:200000 *Caret géologique générale de la Suisse* (*Feuille 6: Sion*). The age of these Flysch sediments south of the Thunersee is generally regarded as Lutetian (Huber, 1933; Cadisch, 1953, p. 174) or early Eocene (Leupold, 1942, p. 284), but systematic investigations are necessary to ascertain their stratigraphic range.

Lithology and Description of Localities

The Leissigen Flysch and the Sandstone Flysch of the Internal Prealps are both similar to the Schlieren Flysch in lithology, but they are on the whole more fine-grained and more calcareous than the Schlieren Flysch, and they contain more flute casts, more slump folds, and fewer groove casts. Directional sedimentary

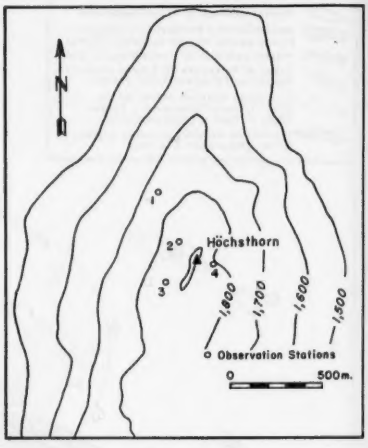

FIGURE 11.—SKETCH MAP SHOWING LOCATION OF OBSERVATION STATIONS IN THE SANDSTONE FLYSCH NEAR ADELBODEN

The Wildstrubel Sheet (Blatt 263), *Landeskarte der Schweiz*, was used as the base map.

structures have been measured from the following localities:

(1) Good exposures of the Leissigen Flysch are present in the Griesbach (Fig. 10, locs. 1, 3) and in a quarry near the Leissigen-Neuenried path at about P. 670 (Fig. 10, loc. 2). The quarry section consists of interbedded sandstone, siltstone, and shale. Some sandstone beds show graded bedding. The siltstone layers are either laminated or cross-laminated. Flute casts were found on the bottom of some sandstone and some siltstone layers. On the steep west bank of the east fork of the Griesbach north of Lammgraben at about P. 730 (Fig. 10, loc. 4) are some Leissigen Flysch sandstone layers that exhibit beautifully oriented large flute casts.

(2) Gurnigel-Schlieren-type Flysch sandstone is well exposed on top of the Höchsthorn south of Adelboden (Fig. 11). On the western slope of the Höchsthorn at about P. 1880 (Fig. 11, loc. 3) is a sandstone bed that exhibits well-preserved flute casts. Sandstone beds containing flute casts are also present on the eastern slope of the Höchsthorn (Fig. 11, loc. 4).

Direction of Transport

Flute-cast orientations indicate that both the Leissigen Flysch and the Sandstone Flysch near Adelboden were deposited by currents that flowed in general from southwest to northeast (Figs. 12, 13). This direction is similar to the easterly to northeasterly direction of transport of the Schlieren Flysch sediments and is the opposite of the southerly direction of transport of the Gurnigel Flysch.

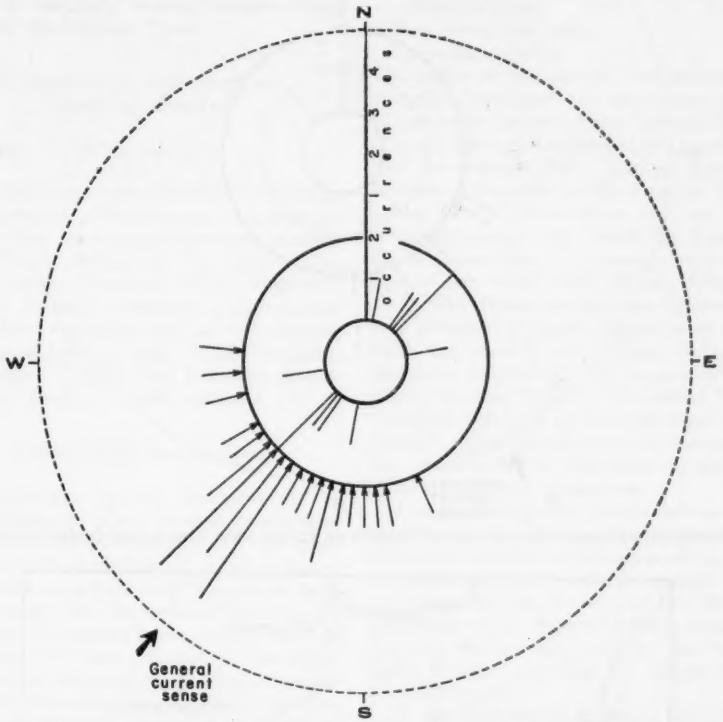

FIGURE 12.—CURRENT-ROSE DIAGRAM, LEISSIGEN FLYSCH, 36 MEASUREMENTS

Paleogeographic Implications

The fact that the general direction of the current depositing the Leissigen Flysch is practically the same as that depositing the Flysch in Sattelzone near Adelboden gives support to previous correlations of the Ultrahelvetic zone south of Thunersee with that of the Internal Prealps. Flysch sediments in these two areas were deposited in the same basin and subsequently overthrust as one tectonic element. They are now separated by erosion and preserved as isolated outcrops above the Helvetic nappes.

Current-structure data shed new light on the tectonic position of the Schlieren Flysch. The rootless Schlieren Flysch has been designated as Ultrahelvetic on the basis of its correlation with the Ultrahelvetic Gurnigel Flysch of the External Prealps and its correlation with the Ultrahelvetic Gurnigel-Schlieren-type Flysch of the Internal Prealps, which stratigraphically overlie older sediments of the Ultrahelvetic facies. Correlation of the Schlieren with the Gurnigel Flysch is now questioned because of differences in flow directions. Correlation of the Schlieren Flysch with the Ultrahelvetic Flysch south of Thunersee is confirmed by the similarity in orientation of their current structures. The Schlieren Flysch, the Leissigen Flysch, and the Ultrahelvetic

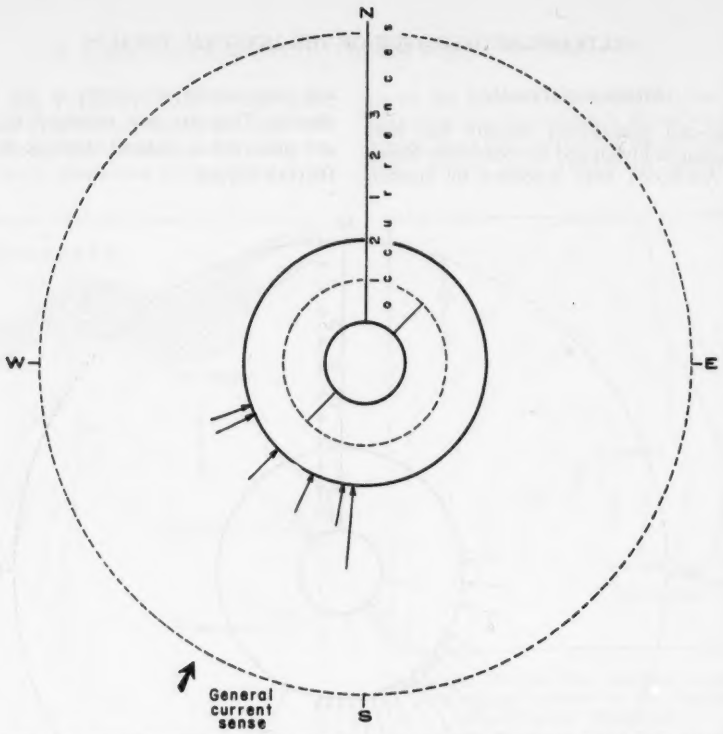

FIGURE 13.—CURRENT-ROSE DIAGRAM, SANDSTONE FLYSCH NEAR ADELBODEN, 8 MEASUREMENTS

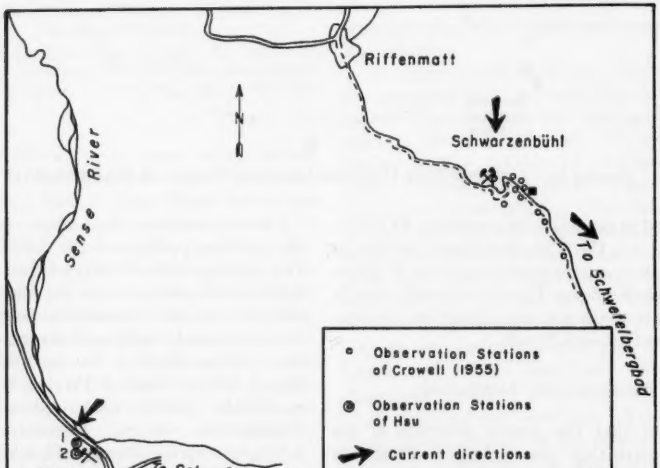

FIGURE 14.—SKETCH MAP SHOWING LOCATION OF OBSERVATION STATIONS IN THE GURNIGEL FLYSCH
The Gantrisch sheet (Blatt 253), *Landeskarte der Schweiz*, was used as the base map.

Flysch in the Sattelzone near Adelboden probably were all deposited in the same basin by northeasterly and easterly flowing currents, which originated from the Habkern massif. This same source area also contributed sediments to southerly flowing currents which deposited the Gurnigel Flysch.

Ultrahelvetic Flysch of the External Prealps

Name and Age

The Sandstone Flysch of the External Prealps in cantons Bern and Freiburg has been generally referred to as the Gurnigel Sandstone (Gerber, 1925; Tercier, 1928a). The Sandstone Flysch of the External Prealps in Waadt (Gagnebin, 1924) is generally considered a stratigraphic and tectonic equivalent and has been mapped as Gurnigel-Schlieren-type Flysch Sandstone in the new 1:200000 *Carte géologique générale de Suisse* (*Feuille 5: Genève-Lausanne*).

Directional-Current Structures

The directional-current structures of the Flysch sandstones of the External Prealps were studied in detail by Crowell (1955). He made more than 100 measurements, which indicated a southerly to southeasterly direction of sediment transport for the Gurnigel Sandstone in canton Bern (Crowell's Schwarzenbühl section). Additional data were obtained by the writer in the nearby Zollhaus Quarry (Fig. 14) on the Plaffeien-Schwarzsee Highway, where the data indicate southwesterly flowing currents (Table 2). Thus, the general sense of the sediment transport of the Gurnigel Flysch is southerly, in contrast to the northeasterly direction of current flows in the Internal Prealps.

Paleogeographic Implications

The paleocurrent information would place the depositional area for the Sandstone Flysch of the Internal Prealps on the north side of a Habkern swell and that of the Gurnigel Flysch on the south side of the same swell. Such a paleogeographic interpretation has a significant bearing on the tectonic position of the nappes of the External Prealps.

Investigations in the Sattelzone (Zone des Cols) by Lugeon (1920), Lugeon and Gagnebin (1941), Gagnebin (1928), Badoux (1945; 1946), and others have led to the following scheme of subdivision of the Ultrahelvetic nappes of the Internal Prealps (Cadisch, 1953):

Meilleret-Decke
Bex-Laubhorn-Decke
Tothorn-Sex Mort-Decke (formerly Mont Bonvin-Decke)
Tour-d'Anzeinde-Decke
Plaine-Morte-Decke

The nappes of the External Prealps have been generally considered the equivalent of the Ultrahelvetic nappes of the Internal Prealps (Fig. 15). Recently Guillaume (1957) considered that the Gurnigel Flysch and its underlying Mesozoic formations of the External Prealps belong to the Plaine-Morte and the Tour-d'Anzeinde-Decke. The Sandstone Flysch in the Sattelzone near Adelboden belongs to the Tothorn-Sex Mort-Decke (Huber, 1933; Cadisch, 1953). Because sediments in higher nappes were deposited in basins situated more to the south than those in lower nappes, Guillaume's correlation implied that the depositional area of the Sandstone Flysch of the Internal Prealps was south of that of the Gurnigel Flysch. This paleogeographic interpretation is contrary to the one presented by the writer on the basis of the paleocurrent information.

If the paleogeographic interpretations based on paleocurrent structures are accepted, two alternative tectonic interpretations of the structure of the External Prealps are possible:

(1) The Gurnigel Flysch and the Mesozoic formations of the External Prealps belong to the same tectonic unit. Together they form a higher tectonic unit than the nappes of the Internal Prealps.

(2) The Mesozoic formations of the External Prealps correlate with the Tour-d'Anzeinde nappe of the Internal Prealps, but the Gurnigel Flysch belongs to a higher tectonic unit.

The stratigraphic evidence seems to dictate that the Mesozoic formations of the External Prealps belong to the Tour-d'Anzeinde nappe[6] and that "On no account can these Mesozoic units come from above the Laubhorn nappe"

[6] Guillaume (1957, p. 169) recognized a weakness in his correlation of the Montsalvens Mesozoic with the Tour d'Anzeinde nappe, because he found no evidence for the Maestrichtian Wang transgression in the External Prealps. The overlap of the late Senonian Wang beds on the bevelled edge of the older rocks is a common characteristic shared by the Ultrahelvetic zone of the Internal Prealps and the southern regions of the Helvetic Alps; if this transgressive relationship is traced, the Ultrahelvetic Tour-d'Anzeinde nappe is found to represent the depositional area just south of the Helvetic zone. (Fig. 17; Leupold, 1942).

(Trümpy, personal communication). Thus, the writer is compelled to accept the second alternative that only the Gurnigel Flysch of the External Prealps represents a higher tectonic element than the nappes of the Internal Prealps.

ently, in late Senonian time, following the deposition of the Senonian Amdener Schichten, and before the transgression of the Wang beds, the Ultrahelvetic realm was mildly deformed and tilted, while more severe deformation—the

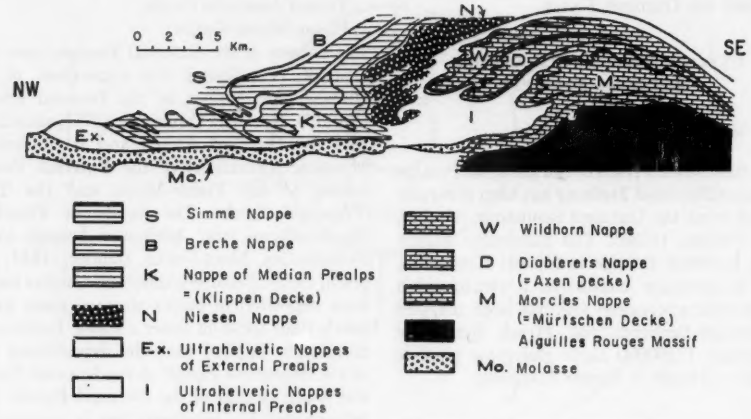

FIGURE 15.—TECTONIC CROSS SECTION THROUGH WESTERN SWITZERLAND
After Gagnebin (1934)

LATE CRETACEOUS-EARLY TERTIARY PALEOGEOGRAPHY

Paleogeographic reconstructions of the Swiss Alps must rely chiefly on stratigraphic and tectonic analyses. However, directional sedimentary structures could yield information on paleocurrent directions. Interpreted within the paleogeographic framework provided by stratigraphic and tectonic studies, such information might prove to be valuable additions to our imperfect geologic record. Additional data would thus permit one to refine previous paleogeographic reconstructions. The following account of the Late Cretaceous-early Tertiary paleogeography of the Ultrahelvetic realm is drawn largely from the work of the various Swiss geologists, particularly Leupold (1942). Several major modifications, however, were necessary in order to accommodate the new paleocurrent information.

Stratigraphic studies (Schaub, 1936; Leupold, 1942; Bolli, 1944; Staeger, 1944) indicated that the Wang beds (primarily Maestrichtian) overlap successively older formation in the southern Helvetic and Ultrahelvetic regions, whereas they are in conformable contact with the underlying Amdener Schichten in the more northerly parts of the Helvetic realm (Fig. 17). Apparently, in late Senonian time, following the deposition of the Senonian Amdener Schichten, and before the transgression of the Wang beds, the Ultrahelvetic realm was mildly deformed and tilted, while more severe deformation—the Niesen phase (Lugeon, 1938)—took place farther south. After subaerial erosion had bared rocks at least as old as Middle Jurassic (Aalenian) in the southern Ultrahelvetic realm, subsidence took place to allow for the transgressive deposition of the marine muds now known as the Wang beds. Either before or during the deposition of these Maestrichtian sediments, the region must have been tilted again so that the deepest part of the trough was located in the more deeply eroded southern part of the region. (*See* Fig. 17.)

In the Glarus Alps of northeastern Switzerland, Leupold (1942) traced a lateral gradation from the Wang beds and the Flysch-like Wang beds in the southern Helvetic realm to the Sardona Flysch in the Ultrahelvetic realm (Fig. 16). He had difficulty, however, in fitting the Wäggital Flysch into his paleogeographic scheme. The sedimentary structure data help solve this uncertainty. Apparently, the Wäggital Flysch was deposited in a basin south of the Sardona basin, and the two were separated by a subaerial barrier that contributed sediments to both (Fig. 16). In central Switzerland, Leupold postulated a land barrier separating the Helvetic realm of Wang beds deposition from the Ultrahelvetic Schlieren Flysch basin. He named the barrier the South Helvetic swell

FIGURE 16.—DIAGRAMMATIC STRATIGRAPHIC SECTION AT EARLY PRIABONIAN TIME, GLARUS ALPS

The Ultrahelvetic Flysch formations (coarse stippled) were probably uplifted during the pre-Priabonian deformation. Meanwhile, subsidence in the Helvetic realm resulted in a transgression of the Priabonian *Globigerina* shale (dashed) over earlier Tertiary carbonates (brick pattern) and quartz sands (fine stippled). The Maestrichtian and the Priabonian Wildflysch are represented by the open circles.

FIGURE 17.—DIAGRAMMATIC STRATIGRAPHIC SECTION AT EARLY PRIABONIAN TIME, CENTRAL SWISS ALPS

The Ultrahelvetic formations (coarse stippled) were probably uplifted during the pre-Priabonian deformation. The South Helvetic swell (SH swell) was also uplifted and was subjected to erosion during this orogenic episode. The Turonian Leimernschichten was exposed by erosion (Leupold, 1942). The South Helvetic swell subsided during the Priabonian epoch, when the *Globigerina* shale (dashed) was deposited directly on the Turonian formation of the swell. The lower Tertiary shore-line sands are represented by the fine stippled pattern, and the lower Tertiary carbonates by the brick pattern. The Priabonian Wildflysch is represented by the open circles.

and postulated that it was the northerly source for the Schlieren Flysch. But the paleocurrent data indicate a southwesterly source for the Schlieren Flysch. Furthermore this South Helvetic swell probably was not active until post-early Eocene time. Thus the writer envisions a gradation from the Helvetic Wang beds to the Ultrahelvetic Schlieren Flysch in central Switzerland similar to that in the Glarus Alps. However, a land barrier existed southwest of the Schlieren basin. This barrier, here designated the Habkern massif, separated the Schlieren basin from the Gurnigel basin farther southwest (Figs. 9, 17).

The land and sea-bottom relief was accentuated during the Paleocene. Much of the Helvetic shelf became dry land, while the Ultrahelvetic Flysch basins were deepened, particularly in central Switzerland where the thick Schlieren sandstone accumulated. The edge of the shelf was the site of shallow-water sedimentation, where the Paleocene sands and lithothamnian limestones were formed (Figs. 9, 16, 17).

The paleogeographic framework was similar during the early and middle Eocene. The shallow sea encroached onto the Helvetic shelf and formed the various units of shore-line sands and limestones, such as the lower Eocene Einsiedelnkalk, the middle Eocene Assilinengrünsand, and Complanatakalk (Fig. 10). Although the lower Eocene shallow-water sediments of northeastern Switzerland were later overthrust as a Schuppen zone under the Ultrahelvetic nappes, similar sediments of the same age situated on the site of the future southern Helvetic swell in central Switzerland were largely eroded away; that they must have been once deposited can only be deduced from the presence of lower Eocene *Foraminifera* as reworked debris in the transgressive Priabonian sediments (Leupold, 1942, p. 283). South of the Helvetic carbonate shelf was the continental slope for pelagic sedimentation where the pre-Priabonian shales of the South Helvetic region were deposited (Brückner, 1952, Fig. 2). Farther south were the deep-water Flysch basins where conditions favorable for intermittent turbidity-current actions resulted in a regular interbedding of turbidity-current-deposited sandstones in pelagic shales. While the shallow sea was gradually encroaching onto the Helvetic shelf during middle Eocene, the Flysch basins were gradually filling up. The margins of some Flysch basins were undergoing deformation; for example, the axis of the Schlieren basin might have shifted northward from the realm of the Laubhorn Decke during Paleocene to the realm of the Tothorn-Sex Mort-Decke during middle Eocene (Fig. 17).

Another major epoch of deformation resulted in a completely changed paleogeographic picture for the upper Eocene. The deep Priabonian sea, in which Globigerina shales were deposited, was restricted to the Helvetic realm and to the Autochthonous zone. The South Helvetic swell was raised in central Switzerland, and its thin mantle of Maestrichtian lower Eocene and middle Eocene sediments was stripped off. Thus, the Priabonian sea transgressed over Turonian in this area (Fig. 17). Later these Turonian rocks were overthrust as the Leimer Schuppen between the main Helvetic and Ultrahelvetic nappes (Leupold, 1942). The Priabonian South Helvetic swell passed westward into the northern branch of the Habkern massif where Habkern-type granite cropped out. Blocks of granite carried westward by submarine slides into the depth of the Priabonian sea formed the Habkern Wildflysch of the Habkern Mulde (Gigon, 1952; Soder, 1949), which is characterized by exotic blocks of the Habkern granite. Farther east, too distant from the source of the Habkern granite, the Habkern Wildflysch contains very few exotic granite blocks (Bentz, 1948; Furrer, 1949; Geiger, 1956).

The information on paleocurrent directions points to the existence of several positive elements in the Ultrahelvetic realm during Mesozoic and early Tertiary time (Fig. 9). The most prominent of these, here designated the Habkern massif, was characterized by a basement composed predominantly of the pink Habkern-type granite. This Habkern massif remained submerged during much of the Mesozoic, though it was uplifted from time to time. Intermittent erosion of its northern margins and repeated transgressions resulted in the various overlapping unconformities observed in the Cretaceous of the Ultrahelvetic and the southernmost Helvetic regions (Staeger, 1944). The Habkern massif became a prominent mountainous peninsula or island during the Maestrichtian-Paleocene time. Northeast and east of the Habkern massif was the Flysch basin of the Internal Prealps and the Schlieren basin. Southeast of the massif were the Flysch basins of the External Prealps. The presence of high mountains adjacent to the deep basins resulted in conditions favorable for turbidity-current deposition. Easterly and northeasterly flowing currents deposited the sandstones of the Schlieren Flysch. Southerly flowing currents

deposited the sandstones of the Gurnigel Flysch (Fig. 9). The Habkern massif was buried under the advancing nappes during late Priabonian time. Today its root possibly lies under the front of the Grand St. Bernhard nappe.

According to these interpretations, the Paleocene-lower Eocene shallow-water sediments of the southern Helvetic realm were deposited in an area on the back of the Gotthard massif. Southeastward and southward from the massif, the Helvetic shelf sediments graded laterally into the Ultrahelvetic Sardona Flysch, in the Glarus Alps and into the Schlieren Flysch in the central Swiss Alps. Separated from these northern Ultrahelvetic Flysch basins by a peninsula range or an island arc were the southern Ultrahelvetic Wäggital and Gurnigel basins (Fig. 9). After the main phase of the Alpine deformation, the southern Helvetic and the Ultrahelvetic realms were pushed northward to form the various units of the Helvetic and Ultrahelvetic nappes (Wildhorn-Drusberg Decke, Wang Schuppen, Leimer Schuppen, Habkern Wildflysch zone with exotic blocks, Schlieren Flysch Decke, and the various Ultrahelvetic nappes of the Internal Prealps and External Prealps).

Present paleogeographic interpretations place the root zone of the uppermost Helvetic nappe at the south slope of the Gotthard massif and the root zone of the Ultrahelvetic Sandstone Flysch farther to the south beneath the present Pennine front. (*Cf.* Figs. 1, 9.) Available stratigraphic and structural information supports such a view. According to Niggli (1944) and Nabholz (1948) the Tavetscher and Gotthard massifs were the underlying basement of the main Helvetic nappes of the Glarus Alps (Cadisch, 1953, p. 334). Slices of crystalline rocks incorporated at the base of a lower Helvetic nappe (Mürtschen nappe) resemble the basement rock type of the Tavetscher massif or that of the northside of the Gotthard massif (Trümpy, 1947). It is thus reasonable to seek the root of the highest Helvetic nappe and that of the Einsiedeln-Blattengrat Schuppen on the south side of the Gotthard massif. Cadisch (1953, p. 355) had this in mind when he noted that "The Mesozoic sedimentary cover on the south side of the Gotthard massif consists in all cases only of Triassic and Liassic rocks. . . . . It is certainly possible that the younger formations were stripped off and thrust northward to form the upper Helvetic and Ultrahelvetic nappes."

If the Gotthard massif is the basement of the higher Helvetic nappes, the apparently anomalous Late Cretaceous-early Tertiary north-south sedimentary strike of the Glarus South Helvetic region (Leupold, 1942), can be readily explained. The sediments of the Einsiedeln-Blattengrat Schuppen zone represent the eastern and southeastern edge of the massif where conditions for shallow-water sedimentation prevailed during Paleocene and early Eocene. The sea deepened eastward and southeastward from the edge of the massif to the Ultrahelvetic Flysch basins (Fig. 9).

Flysch Sedimentation and Paleostructure of the Alpine Geosyncline

The classic ideas on the origin of the Flysch are closely tied to the theory of growing nappes. Schardt (1898) painted a vivid picture of how blocks of granite and sedimentary rocks were peeling off advancing nappes and being dumped into a deep trough filled with muds to form the Wildflysch. Argand (1916) advanced the theory of embryonic nappes. Arbenz (1919) refined Schardt's earlier speculations: Flysch was considered to be orogenic sediment deposited in foredeeps in front of rising and advancing nappes.

The theory of embryonic nappes implies asymmetrical profiles for Alpine basins. The front of the nappe should always form the steep side and the back always the gentle side of a basin. As the Alpine nappes all advanced northward, we should thus expect a gentle southerly dip and a steep northerly dip for the Alpine basins. Flysch sediments, being dumped into the foredeep in front of northward-advancing nappes, should have a southerly source. However, sedimentary structures indicate that the Wäggital Flysch and the Gurnigel Flysch were deposited by southerly flowing currents, presumably derived from northerly sources. Nänny (1946) also suggested a northerly source for the Pennine Prätigau Flysch. The implication of a source area to the north of some Flysch basins demands a reconsideration of the classical idea that the Alpine Flysch represents foredeep deposits of embryonic nappes advancing from the south. Some Flysch sediments may therefore be "backdeep" deposits. The southerly direction of turbidity flow implies a southerly dipping steep slope, possibly resulting from high-angle faulting on the north side of some Flysch basins. This paleogeographic situation suggests that the Alpine geosyncline prior to its final folding was characterized by a

topography typical of block-faulted terranes. Stratigraphic and structural analyses have led to the same idea. Detailed stratigraphic analyses revealed steep southerly dipping slopes of some Alpine basins (Trümpy, 1949). Structural analyses revealed the presence of normal faults during the growth of the Alpine geosyncline (Günzler-Seiffert, 1941). Tectonic syntheses suggested a "horst and graben" structure for the Alpine geosyncline during earlier stages of its development (Staub, 1951). All these evidences make it difficult to deny the important role played by high-angle faulting in shaping the configurations and the submarine topography of the Alpine geosyncline (Trümpy, 1955). Deecke (1912), Tercier (1939), and Crowell (1955) presented evidence to demonstrate that the Flysch sediments were deposited in basins similar to those of the Indonesian Archipelago and to those offshore from the southern California coast. Recent studies in the Carpathians revealed that Flysch basins there had a similar paleostructure before their final folding (Ksiazkiewicz, 1956; 1958; Dzulynski, 1958).

The idea that the Flysch and other Alpine basins might have been created by high-angle faulting does not rule out the probable existence of embryonic nappes, nor is it irreconcilable to Arbenz's idea (1919) that the Flysch represents orogenic sediments. The apparent dilemma might be resolved if we apply a theory for the origin of geosynclines proposed by the writer (1958): that geosynclinal basins were created by crustal faulting soon after folding in order to restore isostatic equilibrium that had been disturbed by anti-isostatic orogenic movements. In the case here considered, the isostatic equilibrium of the Swiss Alps was locally disturbed during pre-Maestrichtian folding (Niesen phase). When the orogenic stress was somewhat relieved to permit isostatic adjustment, Flysch basins were created in places underlain by abnormally thin crust and coastal ranges or island arcs formed in places underlain by abnormally thick crust. A topography typical of block-faulted terranes thus prevailed in the Ultrahelvetic region soon after the folding. Post-Maestrichtian and pre-Priabonian time was locally a period of relative quiescence except for isostatic adjustments. Eroded highlands continued to rise, and sediment-loaded Flysch basins continued to subside until shortly before the Priabonian stage, when another epoch of intense compression again disturbed the isostatic equilibrium. New Flysch basins were created elsewhere (in the northern Helvetic and Autochthonous realms) after this pre-Priabonian deformation. The fact that the Ultrahelvetic Flysch basins were of different ages in different parts of the Alps (Tercier, 1936) might be used as evidence to support Gilluly's theory (1949) that anti-isostatic orogenic movements were local and were not synchronous on a worldwide scale. Ultrahelvetic Flysch basins of the Austrian Alps and of the easternmost Swiss Alps (Vorarlberg Flysch, Fäneren Flysch, Wildhaus Flysch) were created in late Cenomanian or early Turonian time, soon after the intense compressive stress responsible for the pre-Gosau phase of folding of the Austrian Alps[7] had been relieved. This Early Cretaceous episode of deformation was so localized that it did not greatly affect the Ultrahelvetic region of the Central Swiss Alps. There, the isostatic equilibrium was not drastically disturbed until the late Senonian, shortly before the Flysch basins were created.

Isostatic adjustment does not necessarily signify a state of crustal extension. Isostatic adjustment has taken place or is taking place in the Island Arc regions which are under compression today (Ewing and Heezen, 1955; Officer and others, 1957; Vening Meinesz, 1954; Kumagai, 1940; Hsu, 1958, p. 310). Episodes of high-angle faulting do not necessarily represent crustal extension. Isostatic adjustment by high-angle faulting might take place where compressive stress was reduced enough to allow the gravitational field to become the dominant force, resulting in predominantly vertical movements. Thus, isostatic adjustment might take place even when nearby areas are being folded by intense compressive stress.

Umbgrove (1947, p. 62) noted that "The most striking results of an examination of basins and troughs (such as those of the Indonesian Archipelago)[8] is that the origin and history of these formations are related to certain epochs of folding and mountain-building." Such a close time relationship between mountain building and geosynclinal formation has been reviewed by the writer and cited as evidence to support the writer's theory on the origin of geosynclines (Hsu, 1958). Arbenz (1919) also noted a close time relationship between mountain building and Flysch sedimentation. He considered Flysch as orogenic sediments deposited in foredeeps in

[7] For evidence and age of the pre-Gosau phase folding, *see* Schaeffer and others. (1951), Hahn and Zeil (1954), Zeil (1956).

[8] Parenthetical expression inserted by the writer.

front of advancing nappes. The writer believes, however, that sediments of the Flysch facies are postorogenic. They were deposited in deep-sea basins which were underlain by thinner crust and subsided in order to restore the crustal equilibrium that had been disturbed by an anti-isostatic orogeny. The ultimate cause of crustal thinning under Flysch basins can be explained by various speculative hypotheses (Hsu, 1958, p. 306). The writer (1959b) related crustal thinning under Flysch basins to crustal melting near the Mohorovicic discontinuity during an anti-isostatic orogenic episode.

Summary

This study of the Ultrahelvetic Flysch of the Prealps confirms that paleocurrent structures can yield valuable information for paleogeographic and tectonic analyses. The major conclusions of this paper are summarized as follows:

(1) The Ultrahelvetic Flysch basins of the Central Swiss Alps were created as a result of isostatic adjustments subsequent to an episode of pre-Maestrichtian folding. During the Maestrichtian to middle Eocene time, the Ultrahelvetic region was characterized by a topography typical of block-faulted terranes. Rugged land adjoining deep-sea basins resulted in conditions favorable for periodic turbidity-current deposition.

(2) The sandstones of the Wäggital Flysch were deposited by southerly flowing turbidity currents, those of the Schlieren Flysch by easterly to northeasterly flowing turbidity currents, those of the Flysch of the Internal Prealps by northeasterly flowing turbidity currents, those of the Gurnigel Flysch by southwesterly to southeasterly flowing turbidity currents.

(3) The Schlieren Flysch and the Sandstone Flysch of the Internal Prealps were deposited in the same basin. This Schlieren Flysch basin was separated from the Gurnigel Flysch basin by a positive element, herein designated the Habkern massif, which contributed sediments to both Flysch basins and is now buried under the Pennine nappes.

(4) The Sandstone Flysch of the Internal Prealps is generally considered a part of the Tothorn-Sex Mort Decke. If so, the Schlieren Flysch must occupy a tectonic position somewhat equivalent to the Laubhorn Decke, which is practically the highest Ultrahelvetic nappe of the Internal Prealps. The Gurnigel Flysch of the External Prealps has its root in a zone even farther south than the Schlieren Flysch. This origin places the Gurnigel Flysch of the External Prealps as a higher tectonic element than the nappes of the Internal Prealps.

Paleocurrent structures exist in other Alpine Flysch formations. In reconnaissance trips, the writer found flute casts in Simmen Decke Flysch pointing to a northerly source, flute casts and cross-laminations in the Wildflysch of the Habkern zone indicating a westerly source, slump folds in the Vorarlberg Flysch suggesting a northwesterly source, and groove casts in the autochthonous Altdorf Sandstone Flysch of the Glarus Alps showing a southeast-northwest direction of transport. Observations were too few to be significant, but further studies of such current structures might prove useful in better understanding of the paleogeography and tectonics of the Alps.

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Short Notes

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ROLE OF FLUID PRESSURE IN MECHANICS OF OVERTHRUST FAULTING:

DISCUSSION

BY HANS P. LAUBSCHER

Hubbert and Rubey (1959) reject the view of Terzaghi and others that "fluid pressure acting upon the horizontal base of an engineering structure, such as a dam, is effective only over the pore area", and that "the so-called 'uplift pressure' p_u is given by $p_u = f_b p$, where f_b is the 'boundary porosity' of the surface across which the fluid pressure is applied" (p. 135). Reasons for the rejection apparently are empirical and analytical. Empirically, the fact is stressed that "experimental tests . . . made on such materials as loose soils and concrete have led invariably to the result that $f_b \cong 1$" (p. 135). By their analytical treatment the authors arrive at the conclusion that "the pressure p is effective over the entire area of any surface which may be passed through the system irrespective of the fraction of the area lying within the solid-filled part of the space" (p. 133), and that "the pressure p does not terminate at the fluid–solid boundary but is continuous throughout the total space occupied by the fluid and solid combined" (p. 137).

The writer is unable to accept this view for reasons set forth below.

1. The Gauss Theorem

Equation 36, p. 131,

$$\int_V \text{grad } p dV = \oint_A p d\mathbf{A},$$

derived by the authors and regarded by them as a generalized Archimedes principle, is sometimes known as the "general Gauss theorem" in textbooks on vector or tensor analysis (Duschek and Hochrainer, 1950; 1955). Of this the "Gauss theorem" as usually referred to in mathematical physics,

$$\int_V \text{div } \mathbf{E}\, dV = \oint_A \mathbf{E}\, d\mathbf{A},$$

may be shown to be a special case. We prefer the term "Gauss theorem" to "generalized Archimedes principle" in order to stress the fact that it is a geometrical theorem rather than one of hydrostatics, and that to this latter science it may be applied only under some rather restrictive conditions.

According to the authors Terzaghi pointed out that this theorem (and its implication, the Archimedes principle) is derived for the case where pressure acts on a completely closed surface, and that for this reason it is not applicable to porous rocks underground, where any surface of large radius of curvature will intersect solid- and fluid-occupied spaces alternately. The authors then set out to overcome this difficulty by means of an analytical development.

Before examining their analysis in detail it is necessary to make some fundamental remarks about the Gauss theorem. It is a purely geometrical theorem relating surface integral and volume integral in an analytical scalar field. For its validity it is necessary that the surface be closed and that the field be analytical. For its application to hydrostatics it is to be emphasized that the volume integrand grad p is not the gradient of any actual pressure field within the enclosed volume but that of the analytically continued outside field. For instance, the force of buoyancy acting on a tank filled with mercury immersed in water is obtained by integrating the pressure gradient not of the mercury but of the analytically continued field of the surrounding water.

It is important to note further that the surface integral may be used also in cases where the Gauss theorem is inapplicable, that

is, for nonanalytical surface distributions of pressure and for surfaces that are not closed. From the viewpoint of statistical mechanics it portrays the "real" situation by summing up the changes of momentum that are inflicted on the particles impinging on the surface per unit time (*e.g.*, Joos, 1942, p. 487 ff.). The volume integral, on the other hand, has no such physical meaning, as it is the result of a purely mathematical relationship that is valid only under certain conditions as mentioned above.

Extension of the Gauss Theorem to Water-Filled Porous Rocks

For their analytical extension of the Gauss theorem to porous rocks the authors require "only that the pore spaces of the rocks all be interconnected, that they be fluid-filled, and that the pressure p of this fluid and its first directional derivatives vary continuously in the fluid-filled space" (p. 132). Nothing is assumed about the specific character of the fluid, and no conditions are given concerning the nature of the solid—*e.g.*, its elastic properties. Further, it is worthy of note that no use is made in the authors' analysis of the condition that "pore space all be interconnected". All they actually do require is that the pressure in those pores considered parts of pore space belong to the same analytical field.

Since no physical conditions are given for the nature of materials involved, and no laws, by which stress states in the solid are related to those in pore space, the authors' problem still is a purely geometrical one that may be restated as follows: An analytical scalar field is defined for a multiply connected domain. Determine the distribution of values outside the domain of definition. If formulated this way it is obvious that the problem has no solution. Defining a function only for a certain domain implies that outside this domain we are free to choose whatever distribution of values we want, unless this choice is restricted by some additional conditions. Such are not stated for the problem at hand. In general terms, then, the authors set out to obtain by analytical methods a result that cannot possibly follow from the premises.

Specifically, the authors propose to modify slightly their derivation of the Gauss theorem by introducing fractional porosity. They again dissect the body into parallel prisms whose endings now should be alternately in solid- and in fluid-filled space. The force exerted on the solid part of each prism may be computed by adding the forces acting at each fluid–solid interface *and at the bottom and top surfaces* of those prisms that terminate in solid-filled space. These latter forces, however, are not known, as a pressure distribution is defined only for fluid-filled space. We cannot, therefore, proceed with our analysis unless some further assumption is made. This important step, unfortunately, was omitted in the original text but was kindly explained to this writer by the authors. It consists of letting all prisms terminate in fluid-filled space next to their actual endings in solid-filled space. The body comprising these modified prisms (which we may call modification 1) differs by a very small (practically infinitely small) amount from the volume as originally considered, and from this fact it is concluded that the result obtained for the modified body also holds for the original one. However, is this conclusion analytical? Here is the snag: We may consider another modified body in which all the prisms terminate in solid-filled space (modification 2). This body also differs by no more than an infinitesimal amount from the original one, yet the integral of forces acting on the solid frame is finitely different from that of modification 1, *unless we postulate a priori that pressure distribution in solid-filled space be the analytical continuation of pressure distribution in fluid-filled space.* That is, the volume integral will diverge, and integration may not be performed unless this *a priori* assumption is made. The authors, by assuming the volume integral of modification 1 to be the same as that of the original body therefore implicitly assume convergence of the integral and thereby postulate *a priori* that the pressure field of the fluid be continuous throughout solid-filled space.

The Surface Integral

Since there is no *a priori* reason for the volume integral to converge, it is proposed to try simple surface integration. The problem dealt with by the authors is to determine the force of uplift exerted by pore pressure on a body that is to be sheared off its base. Consequently, the surface for which we are to compute the integral is a real or potential shearing fracture. For the computation we may proceed along the lines followed by the authors, with some slight modifications. For the illustration of principles it will be sufficient to deal with a simple case. Let the basal surface of the body

be perfectly plane and horizontal and consider only the vertical component of forces. Let the upper surface of the body be bounded by a nonporous body whose weight constitutes the external load.

We may consider the shearing surface to be composed of two partial surfaces, one consisting of those areas where it passes through solid-filled space (partial surface *a*), and the other of those where it passes through fluid-filled space (partial surface *b*). Perform the integration for the two partial surfaces separately.

First evacuate pore space and integrate over partial surface *a*. The integral must be equal to the external load plus the weight of the solid frame:

$$\int_{A_a}\int \sigma_{zz}\, dA_a = F_s + \rho_s g V(1 - f_v) = F_a .$$

Now fill pore space with fluid. Thereby neither the external load nor the weight of the solid frame is changed. All we do is add the impacts of fluid particles on pore surfaces. From the viewpoint of statistical mechanics the force exerted by a fluid on any obstacle consists of the sum of the changes of momentum per unit time which fluid particles undergo when impinging on the surface of an obstacle. Partial surface *b* therefore consists of the entire surface of interconnected pore space within the body. The latter now may be dissected into vertical prisms as is done by the authors, the prisms extending to the upper surface of the body. The force exerted by the fluid on the solid may be computed for each prism separately, and the total force acting on partial surface *b* is equal to the sum of forces acting on all the prisms. Take one prism and consider a fluid-filled interval that does not terminate at the basal surface. The sum of forces acting on its upper and lower fluid–solid interfaces is directed downward and equal to the weight of the fluid contained in the interval. On the other hand, pore intervals terminating at the basal surface only add momenta directed upward with a total force that equals the product of pressure at the interface times horizontal cross-sectional area of the prism. The sum total of forces acting on partial surface *b* consequently equals the weight of the pore fluid (directed downward) minus the product of pressure at the basal surface times the area of the basal surface times fractional surface porosity of the basal surface. For this formulation we have assumed pressure at the upper surface of basal pore intervals to be essentially equal to that at the basal surface. The total force acting across the basal surface consequently is

$$\begin{aligned} F &= F_a + F_b \\ &= gV(\rho_f f_v + \rho_s(1 - f_v)) + F_s - pAf_b . \end{aligned} \tag{1}$$

That is, the external load plus the total weight (solid and fluid combined) of the body is relieved by pore pressure acting on that part of the shearing surface that passes through pore space. The result is in perfect agreement with Terzaghi's view.

The Terzaghi Experiment

An approach similar to that of the preceding section was chosen by the authors to discuss the Terzaghi experiment (p. 136–137). In essence, this experiment deals with a finite porous body on the external surface as well as on the pore surfaces of which the constant hydrostatic pressure p acts. The weight of the solid and the interstitial fluid is neglected. The result of their analysis is that each point within the solid is under constant hydrostatic pressure p. Exactly the same result is obtained by application of the formulas of the theory of elasticity for a spherical container (Timoshenko, 1934, p. 325–326) when outside and inside pressures are made equal. In fact, if we let the outside wall of the container move to infinity, while keeping the outside pressure equal to the inside pressure, then each point of the practically infinite solid-filled space will be under constant hydrostatic pressure p. Are we not at least in this special case entitled to claim that "the pressure p does not terminate at the fluid–solid boundary but is continuous throughout the total space occupied by the fluid and solid combined" (p. 137)?

To examine this question apply formula (1) of the preceding section. Neglecting the weight of the solid and the interstitial fluid:

$$F = F_s - pAf_b = pA - pAf_b = pA(1 - f_b).$$

The size of partial surface *a*, passing through solid-filled space, is

$$A_a = A(1 - f_b),$$

and there is a compressional stress in the solid of magnitude

$$\sigma_{zz} = pA(1 - f_b)/A(1 - f_b) = p.$$

That is, the authors' result in this case is in perfect agreement with Terzaghi's view and

our own as developed in the preceding section. It is, however, in disagreement with the authors' view stated subsequently (p. 139): "By virtue of the diminution of the exterior stresses by the opposing pressure p, the normal and shear stresses inside the specimen become

$$S' = S - p = \sigma$$

$$T' = T \qquad = \tau".$$

According to this view the compressive force across a shear plane passing through the porous body is zero irrespective of surface porosity. From the preceding analysis it follows that the statement should read instead

$$\sigma = S - pf_b,$$

where $f_b = 1$ from experimental evidence for reasons still to be explained.

Thus, even for this special case the statement that "the pressure p does not terminate at the fluid–solid boundary but is continuous throughout the total space occupied by the fluid and solid combined" needs a qualifying explanation. The action of fluid pressure on a solid body surrounding the fluid is one of tangential tension (cf. Timoshenko, 1934, p. 325–326) and radial compression, that on a body surrounded by the fluid is one of tangential and radial compression. In the Terzaghi experiment the solid is completely surrounded by the fluid. Consequently, although stresses of magnitude p do exist throughout the solid they are compressive and do not "oppose the normal stress".

Effective Porosity Along a Surface of Fracture

Why, then, the invariable experimental result

$$f_b = 1?$$

Although we are in no position to give a final answer to this question it seems to us that a promising approach lies along some such line as sketched forthwith.

In the first place we propose to consider a microscopic model not of static geometry as we did so far, but of one that changes in the course of the experiment. All these experiments terminate in macroscopic shearing displacement, hence the internal geometry *does* change. That the change does not take place abruptly with the initiation of macroscopically measurable movements is evident from numerous treatises on the phenomenon of fracture. Before a fracture spreads a number of holes make their appearance in the continuation of the initial crack. Later on they coalesce to form larger cracks, and finally these combine into one continuous fracture (Irwin, 1958, p. 567 ff.). Of particular interest for this discussion is the following quotation from Irwin (1958):

> "Very small fracture origins in metals and plastics can often be identified on a fracture surface. Microscopic studies of these invariably show an appearance which suggests a rounded cavity rather than a thin crack. Although openings of more crack-like nature such as segments of cleavage may be present both in earlier and later stages of fracture development there are rather general grounds for considering that fracture development also includes openings which more nearly resemble holes than cracks". (p. 567)

This phenomenon is observed by the geologist, too. Landslides first announce themselves by a discontinuous series of cracks; walls of a building coming apart, belts of en echelon fractures or faults in rocks: these all bear evidence to the fact that such developments are ubiquitous. There can be no doubt that the same processes are operative in compressive as well as tensional systems. We therefore conjecture that development of a fracture increases the effective surface porosity until, at the moment macroscopic movements begin, it approaches unity. The stresses necessary to produce the supplementary porosity are those needed to break cohesion. The additional stresses required to overcome internal friction then will be reduced accordingly.

Where the fracture must not be produced first, because of internal discontinuities which were present from the beginning, the problem appears to be much simpler. Such is the case for loose granular materials or for two solid bodies sliding relatively to each other along their external surfaces. All contacts must be point contacts except for those places where radii of curvature are exactly equal and opposite in sign. The statistical probability for such an exception to occur in three dimensions is exceedingly small, except maybe where crystalline cleavage grains form the granular mass, or when two crystalline bodies slide along crystalline cleavage planes. Any "smooth" cut through a non- or polycrystalline body will contain innumerable irregularities on a microscopic scale, as shown by the existence of friction on such surfaces. Therefore, even the contact between apparently smooth bodies with meticulously planed surfaces will consist essentially of a number of point contacts. Perhaps

this is the case, too, for the experiment described by the authors on pages 159–160. Such surfaces, furthermore, should be free from any adsorbed moisture or gas. Quite generally it is to be remarked that the authors' statement that fractional surface porosity equals fractional volume porosity ought to be qualified. The statement is true only for random surfaces in a statistically isotropic porous body that indicate no preference about which particular places of the solid frame they dissect. Fracture surfaces are not of that nature; they follow, within a belt several grain diameters wide, the weakest places, and consequently their surface porosity is greater than the volume porosity to begin with.

Conclusion

From whatever angle the writer looks at the problem of the effect of pore pressure on the mechanics of porous rocks he finds himself in agreement with Terzaghi's view: the force of uplift due to pore pressure is proportional to surface porosity. To explain experimental evidence that along shear fractures the surface porosity is equal to unity he proposes to consider the fact that surface porosity increases during development of the fracture. The Archimedes principle should not be applied to porous rocks underground.

List of Symbols

A_a = partial surface a; that part of the surface passing through solid-filled space

A_b = partial surface b; that part of the surface passing through fluid-filled space.

F_a = force acting on A_a

F_b = force acting on A_b

V = volume of porous body

f_v = fractional volume porosity

f_b = boundary porosity, fractional surface porosity

ρ_s = density of solid

ρ_f = density of fluid

g = acceleration of gravity

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ROLE OF FLUID PRESSURE IN MECHANICS OF OVERTHRUST FAULTING

A Reply

By M. King Hubbert and William W. Rubey

Introduction

In the paper under discussion, the authors were concerned with the mechanical implications of the fact that the rocks in the outer few miles of the lithosphere are to some degree porous and permeable and this pore space is filled with fluids—usually water. The question of what would be the buoyant effect produced by the pressure of this distributed fluid occupying such a three-dimensional interconnected pore network was investigated, and the conclusion reached was that the force of buoyancy would be the same *as if* the fluid pressure were continuous throughout the total space occupied by the fluid and solid combined. Then, if a volume V of this space were to be enclosed by a surface A, the force of buoyancy produced by the pressure p on the total contents—fluid and solid—of this volume would be given by

$$\mathbf{F} = -\iint_A p\, d\mathbf{A}, \tag{1}$$

where $d\mathbf{A}$ is a vectorial representation of the surface element dA, having a direction parallel to the outward normal to dA and a magnitude equal to its area.

This result was found not to be in accord with an earlier inference of Karl Terzaghi (1932; 1936; 1945) according to which the force of buoyancy should be

$$\mathbf{F} = -\iint_A p f_b\, d\mathbf{A}, \tag{2}$$

where f_b is a special kind of porosity called the "boundary porosity" or the "surface porosity" which is the ratio of the pore area to the gross area, along the surface A. In this case the surface A is also assumed to be sinuous so as to pass through the crevices and grain contacts of the solid.

If the pressure p is that of a liquid in hydrostatic equilibrium, then the force of buoyancy $\mathbf{F}$ in equation (1) should be equal to the negative weight of the same volume of the liquid, and should be the same as that which would be given by the principle of Archimedes if the same volume V were to be immersed in a larger space filled with such a liquid. In equation (2), the force of buoyancy would be reduced from that required by the Archimedes principle by the factor f_b. If we distinguish these two buoyancies by the symbols $\mathbf{F}_A$ and $\mathbf{F}_T$, signifying Archimedes buoyancy and Terzaghi buoyancy, respectively, then, if f_b is a constant for the porous system considered, it will be seen from equations (1) and (2) that

$$\mathbf{F}_T = f_b \mathbf{F}_A . \tag{3}$$

All known experiments, including those by Terzaghi, give the result:

$$\mathbf{F} = \mathbf{F}_A , \tag{4}$$

in the light of which the validity of equation (2) can be maintained only by postulating that the surface porosity of all porous solids tested, ranging from loose soils to solid rocks and concrete, has the value of unity.

It was the authors' conclusion that equation (2) is invalid; whereas the critic, in the foregoing discussion, maintains that equation (2) is correct and equation (1) is invalid. His argument is divided into two or perhaps three principal parts, the first negative, the second positive, and a third which is an elaboration of the second.

Negative Part of Critic's Discussion

The negative part consists in trying to show that the authors' analysis of buoyancy in the original paper, on pages 130–137, is not valid. The critic first objects to the authors' derivation of the Archimedes principle as given on pages 130 and 131, where it was shown that the force of buoyancy exerted by a fluid on a completely immersed, isolated solid is given by either of the equations

$$\left.\begin{aligned} \mathbf{F} &= -\iint_A p\, d\mathbf{A}, \\ \mathbf{F} &= -\iiint_V \operatorname{grad} p\, dV, \end{aligned}\right\} \qquad (36, \text{p. } 131)$$

on the grounds that p is not defined in the space occupied by the solid.

This point may not have been made sufficiently clear in the original paper, but the pressure p in the space occupied by the solid is that of the displaced fluid. In all cases this was plainly implied by the equation

$$\operatorname{grad} p = \rho_f \mathbf{g}. \qquad (31, \text{p. } 131)$$

Hence, the critic's remarks with regard to this point appear to be irrelevant, since in no instance was grad p for the completely immersed solid given any other interpretation than that above.

However, for the completely immersed solid, even this restriction is not necessary because if $p'(x, y, z)$ is any pressure field which varies continuously and has continuous first derivatives inside the volume V and coincides with the actual pressure p of the fluid upon its exterior surface A, then in this case also

$$\iiint_V \operatorname{grad} p'\, dV = \iint_A p\, d\mathbf{A}, \qquad (5)$$

and the force of buoyancy is likewise given by the volume integral

$$\mathbf{F} = -\iiint_V \operatorname{grad} p'\, dV. \qquad (6)$$

Hence, so long as the conditions specified above are satisfied, the value of the volume integral does not depend upon any particular definition of the pressure p inside the volume V but only upon the pressure of the actual fluid at the boundary.

Extension to fluid-filled porous solids.—In extending their results obtained for the completely immersed solid to the case of the fluid-filled porous solid, the authors' statement at the outset of the problem they proposed to consider was the following:

> "*Extension of Archimedes principle to water-filled porous rocks.*—Let us now consider whether the principle of Archimedes can be extended to include the case of the porous water-filled rocks of present interest. In this case let us require only that the pore spaces of the rocks all be interconnected, that they be fluid-filled, and that the pressure p of this fluid and its first directional derivatives vary continuously in the fluid-filled space."

Since the solid is specified to be porous rock, and the fluid, water, it follows that the statement of the problem implies all of the mechanical properties, such as density and elastic constants, possessed by rocks and water. Furthermore, the stipulation that the pores be

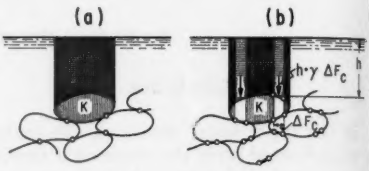

FIGURE 1.—TERZAGHI'S FIGURE 8 (1932) DEPICTING BUOYANT FORCES ON BEDS OF SEDIMENT

(a) with point grain contacts; (b) with finite areas of contact

interconnected assures the existence of a pressure field that will be macroscopically continuous throughout the space considered. However, concerning this question, the critic asserts:

> "Nothing is assumed about the specific character of the fluid, and no conditions are given concerning the nature of the solid, e.g., its elastic properties. Further, it is worthy of note that no use is made in the authors' analysis of the condition that 'pore space all be interconnected.' "

He then dismisses the physical and geological formulation of the problem given by the authors and rephrases it into the following nonequivalent mathematical abstraction devoid of all physical content:

> "An analytical scalar field is defined for a multiply connected domain. Determine the distribution of values outside the domain of definition."

Using this abstraction as a straw man, he then proceeds to demolish it with the following statement:

> "If formulated this way it is obvious that the problem has no solution. . . . In general terms, then the authors set out to obtain by analytical methods a result that cannot possibly follow from the premises."

Positive Part of Critic's Discussion

Since all the remaining negative part of the critic's discussion is based upon the logical hiatus introduced in the statements just quoted, it would be time consuming and profitless to consider this negative part of his discussion in

any further detail. More informative, and of greater importance, is the positive part of his discussion wherein he sets up a concrete problem to which he then gives his own demonstration of the validity of the Terzaghi theory.

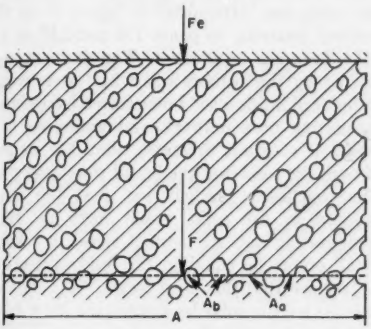

FIGURE 2.—ILLUSTRATION OF SYSTEM ANALYZED BY CRITIC

Before taking this up, however, let us refresh our memories as to just what the Terzaghi theory implies. This theory had its origin in a paper published in 1932 entitled *Tragfähigkeit der Flachgründungen* in which there occurs on pages 666–667 a section on "Hydrostatischer Auftrieb im Ton" ("Hydrostatic uplift [or buoyancy] in clay"). In Figure 8 of this paper (here reproduced as Fig. 1) there are shown two separate beds of sediment beneath static bodies of water: (a) a bed of sand or gravel in which the grain contacts are indicated as being essentially points, and (b) a bed of another sediment, presumed to be typical of a clay, wherein the areas of contact between the grains represent a significant fraction of their total exterior surfaces.

Regarding this problem, Terzaghi reasoned that for the grains which had negligible areas of contact, complete Archimedes buoyancy could be expected, but for the sediment whose grains had finite areas of contact, the pressure of the water would be effective only over the exposed parts of the grains, and would be blanked out over the areas of contact. He concluded, accordingly, that in this case the complete Archimedes buoyancy could not prevail and would in fact be reduced by the factor $(1 - F_c)$, where F_c is the horizontal projection of the area of grain contact per unit area of a nearly horizontal surface passed through the sediment so as to coincide with these contact surfaces.

The factor $(1 - F_c)$ in this formulation is the same as the surface porosity f_b which was later used (with the symbol n_b) by Terzaghi (1945, p. 784) and which leads to the formulation of buoyancy as expressed by equations (2) and (3) herein.

The problem that the critic has set up is represented graphically in Figure 2. Here a horizontal impervious stratum overlies a porous stratum (of which, incidentally, no elastic properties are given) in which a horizontal potential slip surface is to be considered. A portion of this surface of area A is taken as the base of a vertical column. The surface A is divided into the two partial surfaces A_a and A_b, the first being that fraction of A which lies within the solid part of the porous space, and the second that part of A which lies within the pore space.

In this vertical column the weight of the overburden, or the external force, which is applied downward on the top of the porous section by the impervious layer above is stated to be the fixed force F_e. The solid part of the porous body has a density ρ_s and a volume porosity f_v. The surface porosity of the area A is f_b, whereby the partial surfaces A_a and A_b are related to the total surface A by

$$\left.\begin{aligned} A_a &= (1 - f_b)A, \\ A_b &= f_b A. \end{aligned}\right\} \tag{7}$$

The critic wishes to find the total downward force F which is exerted upon the area A, and to do this he undertakes to find the partial forces F_a and F_b exerted upon the areas A_a and A_b separately. To find the force F_a he lets the porous system be evacuated and obtains the correct result that

$$F_a = \iint_{A_a} \sigma'_{zz}\, dA_a = \rho_s(1 - f_v)gV + F_e,$$

(critic's unnumbered equation)

which is simply a statement that F_a is the sum of the external force F_e and the weight of the drained porous solid above A. (The notation has here been used instead of the critic's σ_{zz} σ'_{zz} to distinguish this from the authors' σ_{zz} which has a different meaning).

He next lets the pore space be filled with a fluid. By this operation, he argues, neither the external load nor the weight of the solid frame is changed, so that the partial force F_a, as given by the equation above, would not be affected. Thus it remains necessary only to find the additional partial force F_b induced by the

presence of the fluid, whereby the total force will be

$$F = F_a + F_b.$$

At this point, he redefines the area A_b by the statement: "Partial surface b therefore consists of the entire surface of interconnected pore space within the body." It will be noted that this differs significantly from his original definition of the surface A_b as being that portion of the horizontal surface A "where it passes through fluid-filled space (partial surface b)." Then, on the basis of this redefinition, he evaluates the integral of the pressure exerted by the fluid on the solid over the entire fluid–solid interface in the space above the surface A and arrives at the result:

$$F_b = \rho_f g V f_v - p f_b A.$$

Adding this to F_a as already determined, he obtains the final result that the total downward force exerted by the overburden on the surface A is

$$F = F_a + F_b = gV[\rho_f f_v + \rho_s(1 - f_v)] + F_e - pAf_b. \qquad \text{(Critic's eq. 1)}$$

Concerning this he states:

"That is, the external load plus the total weight (solid and fluid combined) of the body is relieved by pore pressure acting on that part of the shearing surface that passes through pore space. *The result is in perfect agreement with Terzaghi's view.*" [Italics ours]

That the result is also erroneous can be seen by inspection. The total force F acting downward on the surface A must be the weight of the total overburden, or

$$F = F_s + W_s + W_f, \qquad (8)$$

where

$$\left.\begin{aligned} W_s &= \rho_s g(1 - f_v)V, \\ W_f &= \rho_f g f_v V, \end{aligned}\right\} \qquad (9)$$

and

are the weights, respectively, of the solid and fluid fractions of the contents of volume V. Substituting equations (9) into (8) then gives the correct answer:

$$F = F_e + [\rho_f f_v + \rho_s(1 - f_v)]gV. \qquad (10)$$

When this is compared with the critic's equation (1), it will be seen that the latter is in error by the amount of the term $(-pAf_b)$—that is to say, the boundary-porosity term. Then, when it is considered further that p may have any value up to pressures capable of floating the overburden, and f_b any value up to unity, it will be seen that the magnitude of this error may be as great as the total weight of the overburden.

Nevertheless, having arrived at this result the critic has attempted to apply it to the authors' analysis, on pages 136 and 137, of the 1932 Terzaghi experiment. The problem considered by the authors was what stress would be induced inside a solid penetrated by an interconnecting pore network and immersed in a liquid at constant pressure, when the weights of both the solid and the liquid were neglected. The conclusion reached was that the stress induced in the solid would have no shear components, and the normal stress, σ', would at all points be equal to the fluid pressure p.

To show that this must be so, it is necessary only to let the solid be intersected by a plane and then to write the equation of equilibrium of the forces acting on the mass lying on one side of the plane. Considering for the moment only the force components normal to the intersecting plane, and using the present notation, the equation of equilibrium will be

$$F - (F_a + F_b) = 0, \qquad (11)$$

where F is the force component normal to the intersecting plane exerted upon the solid and its fluid contents by the fluid pressure acting over its exterior surface, and $-(F_a + F_b)$ is the force exerted across the area of intersection A. As heretofore, the forces F_a and F_b are the partial forces exerted across the solid and fluid fractions, A_a and A_b, respectively, of the total surface A.

We have, therefore,

$$\left.\begin{aligned} F &= pA, \\ F_a &= \sigma'_{zz}A_a = \sigma'_{zz}(1 - f_b)A, \\ F_b &= pA_b = pf_bA. \end{aligned}\right\} \qquad (12)$$

Then when we substitute equations (12) into (11) and eliminate A, we obtain

$$p(1 - f_b) = \sigma'_{zz}(1 - f_b)$$

from which

$$\sigma'_{zz} = p. \qquad (13)$$

This is the authors' original result as given in equation (65) on page 137.

The critic has attempted to show that the same result can be obtained from his equation (1) and that, therefore, the two analyses are equivalent. When the weights of the solid and

liquid are neglected, he shows that his equation (1) simplifies to

$$F = F_a - pAf_b = pA - pAf_b = pA(1 - f_b).$$

This he then equates to

$$F_a = \sigma'_{aa}(1 - f_b)A$$

and obtains

$$pA(1 - f_b) = \sigma'_{aa}(1 - f_b)A$$

from which

$$\sigma'_{aa} = p,$$

in agreement with equation (13), above.

Concerning this he remarks: "That is, the authors' result in this case is in perfect agreement with Terzaghi's view and our own as developed in the preceding section."

This perfect agreement, however, has been obtained at the expense of two compensating errors: the first arising from the use of the critic's equation (1) which already has been shown to be erroneous, and the second due to the equating of F to F_a when, by the critic's original definition,

$$F = F_a + F_b,$$

and in the problem considered F_b is not zero. Had F_b not been thus omitted from the analysis, the result would have been

$$pA(1 - f_b) = \sigma'_{aa}(1 - f_b)A + pf_bA,$$

from which

$$\sigma'_{aa} = \frac{1 - 2f_b}{1 - f_b} \cdot p.$$

This, however, does not agree with the authors' results; neither is it correct. In this case the error arises entirely from the use of the critic's equation (1).

When the critic's analysis of the problem of his own choosing is studied in detail, it is found that his errors result in part from a fundamental fallacy of assumption, namely that the stress σ' inside the solid is uninfluenced by the pressure of the fluid in the pore spaces. This assumption was involved in his derivation of his equation (1), where the expression for F_a was first derived when the porous solid was drained and the pressure was zero, and then the value so obtained was assumed to be still valid when the porous system was filled with fluid having a finite pressure p at the level of the surface A.

The same assumption is the basis for most of his earlier criticisms of the authors' analysis, and it is also implicit in Terzaghi's (1932) analysis of his clay-cake experiment. In that case the pressure p was simply blanked out in the areas of grain contact without recognizing that a stress of the same magnitude had been induced inside the solid by the fluid pressure in the pore spaces.

Surface Porosity Along a Surface of Fracture

The third part of the critic's discussion is devoted to trying to justify the assumption that the surface porosity f_b along a surface of fracture must always be essentially unity. However, his argument in this case is based upon the premise that his equation (1) is valid. Since this premise has been shown not to be acceptable, no further consideration of that part of the discussion appears to be necessary.

Restatement of the Buoyancy Problem

Although the critic has not demonstrated, as he set out to do, that the authors' view of the buoyancy problem as expressed by equation (1), herein, is erroneous, or that the contrary view, expressed by equation (2), is the correct one, the problem nevertheless is one of fundamental importance. Moreover the authors' treatment of it may not have been sufficiently clear to be convincing. They should like, therefore, to present a brief restatement of the problem and to submit certain additional supporting evidence.

In a fluid-filled porous solid having interconnecting pores, two macroscopic stress systems exist, and are in principle capable of measurement. These are: the total stress and the fluid pressure. If a macroscopic view is taken, where the lengths, areas, and volumes considered are large as compared with the grain and pore dimensions of the system, and if a force ΔF, having normal and tangential components ΔF_n and ΔF_t, is exerted across a small macroscopic plane area ΔA inside the system, the normal and tangential components of the total stress field across ΔA will be given by

$$\left.\begin{aligned} S &= \Delta F_n/\Delta A, \\ T &= \Delta F_t/\Delta A. \end{aligned}\right\} \qquad (14)$$

The total tensor of this stress field at a given point in the system, that is, its components

with respect to the x-, y-, and z-axes, will then be given by

$$\Sigma \equiv \begin{vmatrix} S_{xx} & T_{xy} & T_{xz} \\ T_{yx} & S_{yy} & T_{yz} \\ T_{zx} & T_{zy} & S_{zz} \end{vmatrix}. \tag{15}$$

Likewise, if the fluid pressure is measured at a given point in this system by means of a manometer whose terminus is large compared with the grain and pore size, a definite pressure p will be observed, and the totality of all such measurements throughout the system will comprise a continuous scalar field in terms of the pressure p. Through this field continuous surfaces of equal pressure can be drawn, and at any point a definite value of grad p may be determined.

Regarded as a stress, the pressure p, under conditions which prevail underground, may be taken to be a tensor whose normal components differ negligibly from one another, and whose shear components are essentially zero. This pressure at any given point may thus be represented in the tensor form by

$$\Pi \equiv \begin{vmatrix} p & 0 & 0 \\ 0 & p & 0 \\ 0 & 0 & p \end{vmatrix}. \tag{16}$$

Since the Π-field is only a partial stress and is a component of the total stress, Σ, then it follows that a second partial stress is definable as the difference between Σ and Π, or by

$$\Omega = \Sigma - \Pi. \tag{17}$$

The components of the Ω-field are accordingly given by

$$\begin{aligned} \Omega &\equiv \begin{vmatrix} \sigma_{xx} & \tau_{xy} & \tau_{xz} \\ \tau_{yx} & \sigma_{yy} & \tau_{yz} \\ \tau_{zx} & \tau_{zy} & \sigma_{zz} \end{vmatrix} \\ &= \begin{vmatrix} (S_{xx} - p) & T_{xy} & T_{xz} \\ T_{yx} & (S_{yy} - p) & T_{yz} \\ T_{zx} & T_{zy} & (S_{zz} - p) \end{vmatrix}, \end{aligned} \tag{18}$$

from which, by comparison of corresponding components, it will be seen that for a plane of any orientation

$$\left.\begin{aligned} \sigma &= S - p, \\ \tau &= T. \end{aligned}\right\} \tag{19}$$

It is to be emphasized that by its definition, each of these three stress fields is to be regarded as being continuous throughout the space considered, and quite independent of such microscopic considerations as the surface or volumetric porosity of the system.

Some of the properties of the two partial fields, Π and Ω, can be inferred by inspection. Since the Π-field has zero or negligible shear components, then it is incapable of producing any kind of distortion or change of shape in the porous solid; it can induce a volume change only in accordance with the volume coefficient of elasticity. But this is a second-order quantity as compared with the macroscopic volume changes which can occur in a porous solid when it undergoes compaction and so may be considered negligible for present purposes.

The second partial stress field, Ω, retains all the macroscopic shear components of the total field, and in addition, even if its principal stresses σ_1, σ_2, and σ_3 were all equal and its macroscopic shear components were zero, it would still induce microscopic shear stresses which would cause distortion in the individual grains.

Consequently, following the usage proposed by Terzaghi (1936, p. 874), we may refer to the Σ-field as the *total* stress field, the Π-field as the *neutral* stress field, and the Ω-field as the *effective* stress field.

It is worthy of note at this point that in Terzaghi's (1932) clay-cake experiment, where in a metal ring filled with wet clay was subjected to a large hydrostatic pressure, the partial stress Π and the total stress Σ (if we neglect the very small stresses resulting from the weight of the clay) were identical. Consequently the effective stress Ω must have been essentially zero. Therefore, contrary to Terzaghi's expectations, the cake should not have become compacted simply because no shear stresses were induced. Hence the failure of the cake to become compacted conveys no information whatever concerning the magnitude of either the surface or volumetric porosity of the clay.

With regard to the buoyancy problem, we may now state, quite unequivocally, that if volume V, in a system in which the stress field Π is determinate or definable, is enclosed by a smooth surface A, the force exerted upon the contents of this volume by this field is given by either of the two equivalent equations

$$\left.\begin{aligned} \mathbf{F}_\Pi &= -\iint_A p \, d\mathbf{A}, \\ \mathbf{F}_\Pi &= -\iiint_V \operatorname{grad} p \, dV, \end{aligned}\right\} \tag{20}$$

which involve in no manner either the surface or the volumetric porosity. Then if $\mathbf{F}_\Omega$ is the force exerted upon the same volume by the effective stress Ω, and m is the mass contained within V, the complete equation of translational equilibrium is given by

$$\mathbf{F}_\Omega + \mathbf{F}_\Pi + m\mathbf{g} = 0$$

or

$$\mathbf{F}_\Omega = -(\mathbf{F}_\Pi + m\mathbf{g}). \tag{21}$$

Then introducing the values of $\mathbf{F}_\Pi$ from equations (20) and noting that

$$m\mathbf{g} = \iiint_V \rho_b \mathbf{g}\, dV, \tag{22}$$

where ρ_b is the bulk density of composite mass, we obtain

$$\mathbf{F}_\Omega = \iiint_V \text{grad } p\, dV - \iiint_V \rho_b \mathbf{g}\, dV$$
$$= -\iiint_V (\rho_b \mathbf{g} - \text{grad } p)\, dV. \tag{23}$$

Then, in the special case where p is the pressure of a liquid of density ρ_l in hydrostatic equilibrium,

$$\text{grad } p = \rho_l \mathbf{g}, \tag{24}$$

and equations (20) and (23) become, respectively,

$$\mathbf{F}_\Pi = -\rho_l \mathbf{g} V, \tag{25}$$

$$\mathbf{F}_\Omega = -\iiint_V (\rho_b - \rho_l)\, \mathbf{g}\, dV. \tag{26}$$

Finally, if ρ_b is also constant, equation (26) simplifies still further to

$$\mathbf{F}_\Omega = -(\rho_b - \rho_l)\mathbf{g}V. \tag{27}$$

It will accordingly be seen that the stress field Π extending through a liquid-filled porous solid exerts upon a circumscribed volume a force $\mathbf{F}_\Pi$ which, when the liquid is static, is exactly the same as the Archimedes force of buoyancy that would be exerted upon the same volume if it were isolated and completely immersed in a liquid of density ρ_l. The complementary stress field Ω exerts upon the same volume a force $\mathbf{F}_\Omega$ which is the same as the negative of the submerged weight of the volume when immersed in the liquid of density ρ_l.

Experimental Verification

Since equations (1) and (2),

$$\mathbf{F} = -\iint_A p\, d\mathbf{A} \tag{1}$$

$$\mathbf{F} = -\iint_A f_b\, p\, d\mathbf{A}, \tag{2}$$

differ only by the surface-porosity factor f_b, it should be a simple matter to devise an experiment for which the integrals to the right in the two equations differ from one another, and at the same time permit $\mathbf{F}$ to be measured directly and thus determine experimentally which of the two equations is correct. Such an experiment could be performed with the apparatus shown in Figure 3. This consists of a tank filled with a liquid of density ρ_l in the bottom of which are placed four identical submerged spring balances. Each balance consists of a vertical solid piston which is free to move without friction through a cylindrical sleeve in the bottom of the tank, and is supported by a coiled spring in the air chamber underneath.

Attached to the top of each piston is a pointer opposite a fixed vertical scale. Consequently the equilibrium position for the submerged but otherwise unloaded piston will define the null position on the scale. The balance can then be calibrated by measuring the forces required to produce various deflections, upward and downward, of the piston from this position.

Suppose now that balance (*a*) be left as a control while balances (*b*), (*c*), and (*d*) are loaded with three identical cylinders having the same diameter as that of the pistons of the balances. However, let the densities of the load cylinders be, respectively, $\rho_s = \rho_l$, $\rho_s < \rho_l$, and $\rho_s > \rho_l$. Also let the load cylinders be thoroughly cemented to the balance pistons so as to exclude completely the liquid from the contact and thus produce a surface porosity at the contact of $\rho_b = 0$.

Having placed the three load cylinders on their respective spring balances, let us now compute the force $\mathbf{F}'$ which each of these will exert upon the balance tending to move it from its null position of equilibrium. In each instance the force measured by the balance will be given by

$$\mathbf{F}' = m\mathbf{g} + \mathbf{F} = \rho_s \mathbf{g} V + \mathbf{F}, \tag{28}$$

where m is the mass of the load cylinder, V its volume, and $\mathbf{F}$ is the force of buoyancy which may be calculated by means of either equation (1) or equation (2), whichever is assumed to be valid.

Assuming that $\mathbf{F}$ is correctly given by equation (1), then throughout the space beneath

the surface of the liquid we may set

$$\left.\begin{aligned} p &= \rho_l g z, \\ \text{grad } p &= \rho_l \mathbf{g}, \end{aligned}\right\} \tag{29}$$

In this, the integral over the base vanishes because $f_b = 0$; that over the sides vanishes because the pressures are equal and opposite; so we are left only with the integral over the

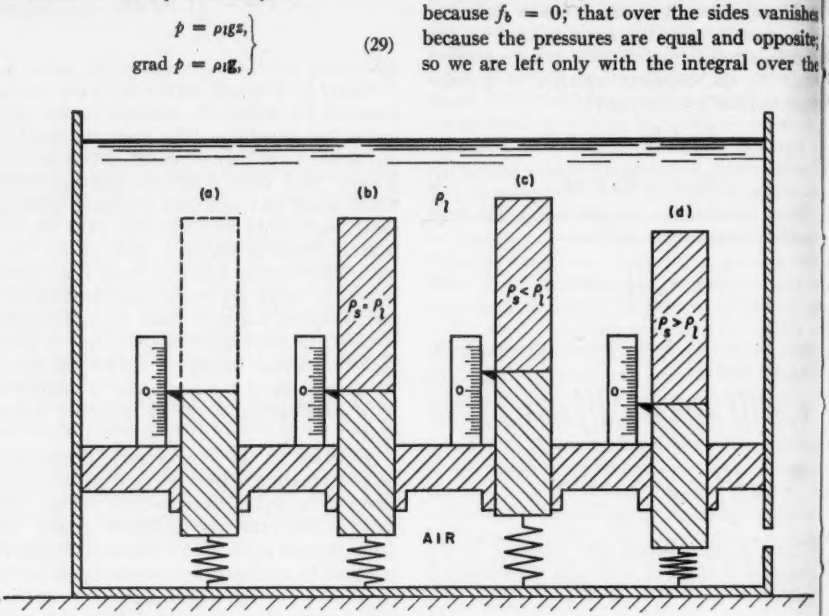

FIGURE 3.—LIQUID-FILLED TANK WITH SUBMERGED SPRING BALANCES FOR MEASURING FORCES OF BUOYANCY ON SOLID CYLINDERS FROM WHOSE BASES FLUID IS EXCLUDED

where z is measured from the liquid surface and taken to be positive in the downward direction. Then

$$\begin{aligned} \mathbf{F}_1 &= -\iint_A p \, d\mathbf{A} \\ &= -\iiint_V \text{grad } p \, dV = -\rho_l \mathbf{g} V, \end{aligned} \tag{30}$$

which when substituted into equation (28) gives for the submerged weight of the cylinder,

$$\mathbf{F}_1 = (\rho_s - \rho_l)\mathbf{g}V. \tag{31}$$

If, on the other hand, we assume equation (2) to be valid, then the buoyancy force will be

$$\begin{aligned} \mathbf{F}_2 = -\iint_A f_b p \, d\mathbf{A} &= -\iint_{\text{base}} f_b p \, d\mathbf{A} \\ &\quad - \iint_{\text{sides}} f_b p \, d\mathbf{A} - \iint_{\text{top}} f_b p \, d\mathbf{A}. \end{aligned} \tag{32}$$

top of the cylinder. Here, $f_b = 1$, and, if the depth of the top of the cylinder is z_1,

$$\begin{aligned} \mathbf{F}_2 = -\iint_{\text{top}} f_b p \, d\mathbf{A} &= -\rho_l g z_1 \mathbf{A}_{zz} \\ &= +\rho_l \mathbf{g} V_l, \end{aligned} \tag{33}$$

where V_l is the volume of the column of liquid above the solid column. When this is substituted into equation (28) we obtain for the net force exerted upon the spring balance,

$$\mathbf{F}'_2 = \rho_s \mathbf{g} V + \rho_l \mathbf{g} V_l. \tag{34}$$

If we now compare equations (31) and (34) we shall see that according to equation (31) the force $\mathbf{F}'_1$ should be zero when $\rho_s = \rho_l$, upward when $\rho_s < \rho_l$, and downward when $\rho_s > \rho_l$, as is shown in Figure 3. According to equation (34), on the contrary, the force applied to the spring balance, $\mathbf{F}'_2$, should be downward in all three cases and equal to the combined weights of the solid cylinder and the superposed column of liquid.

To determine whether equation (1) of

equation (2) gives the correct value of the buoyancy, it is not necessary actually to perform the experiment shown in Figure 3 since it is so simple that the answer can be obtained from elementary mechanics by inspection. Consider balance (*b*) which has been loaded with a cylinder of density $\rho_s = \rho_l$. According to equations (1) and (31), the balance should have no deflection, whereas, according to equations (2) and (34), the deflection should be downward.

However, if we compare balance (*b*) with

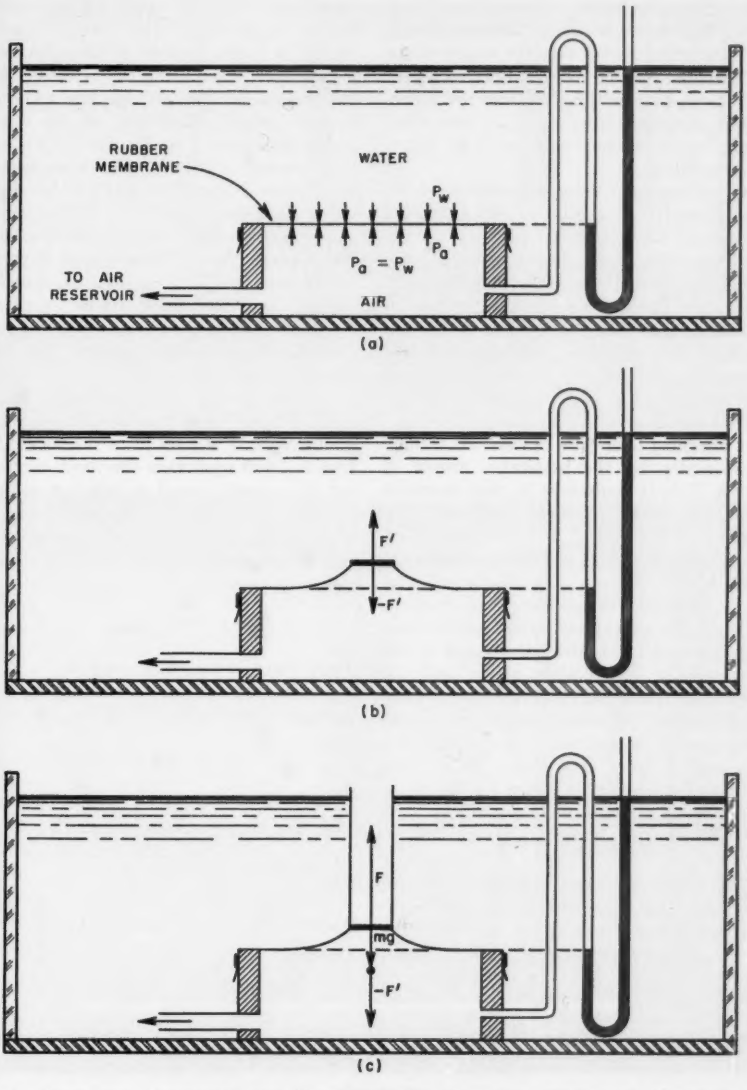

FIGURE 4.—SIMPLIFIED APPARATUS FOR DEMONSTRATING BUOYANCY EFFECT ON CYLINDER FROM WHOSE BASE FLUID IS EXCLUDED

balance (*a*) it will be seen that the placing of the load cylinder on (*b*) has merely resulted in the displacement of an equal volume (and mass) of liquid, and the total load on each balance is exactly the same. Therefore, despite the exclusion of the fluid pressure from its base, the net load added to the balance is zero, and the deflection should be zero. In general, the net load on each of the balances, which will be the difference between the weight of the added cylinder and that of the displaced liquid, will be given by equation (31), and the deflection will be upward when $\rho_s < \rho_l$, zero when $\rho_s = \rho_l$, and downward when $\rho_s > \rho_l$, as indicated in Figure 3.

That this result is quite independent of the surface porosity can also be seen by engraving a pattern of interconnected porosity on the base of the load cylinder. In this manner f_b can be given any value between the limits 0 and 1, and the fluid pressure admitted to the base of the column accordingly. However, in a thin horizontal pore space the fluid pressure will exert equal and opposite forces upon the upper and lower surfaces of the pore lamina, and the net effect will be zero.

The foregoing inferences can also be verified experimentally by the apparatus shown in Figure 4. This is equivalent to that shown in Figure 3 but somewhat easier to construct. In this case the submerged spring balance is an air cell consisting of a rubber membrane stretched over the top of a plastic cylinder which is cemented to the bottom of the tank (Fig. 4 a). The null configuration occurs when the air pressure inside the cell is equal to the liquid pressure immediately above and the membrane is horizontal. If an additional force $\mathbf{F}'$ is applied to the membrane, it will be deflected upward or downward in the direction of the applied force (Fig. 4 b). Then if the membrane is loaded, as in Figure 4 c, the equation for the equilibrium of forces will be

$$(-\mathbf{F}') + m\mathbf{g} + \mathbf{F} = 0, \tag{35}$$

where $\mathbf{F}'$ is the submerged weight of the load as measured by the deflection of the membrane, $m\mathbf{g}$ is its weight in air, and $\mathbf{F}$ is the force of buoyancy exerted upon the load. Thus, by measurement of $\mathbf{F}'$ and $m\mathbf{g}$, the force of buoyancy $\mathbf{F}$ can be determined.

In Plates 1 and 2 photographs of the apparatus are shown in which the load consists of a light-weight plastic hollow cylinder inside which different masses may be placed. The base of the cylinder is firmly cemented to the membrane, and its open top extends just above the surface of the water. Figure 1 of Plate 1 is a general view of the apparatus. In Figure 2 of Plate 1 the cell is loaded with a solid metal cylinder so that the mass of the load is greater than that of the displaced water. In Figure 1 of Plate 2 the cylinder is filled almost to its top with water so that its total mass is equal to that of the displaced water. In Figure 2 of Plate 2 the cylinder is empty, and its mass is much less than that of the displaced water. From the photographs it will be seen that in these three cases the membrane is depressed below its null position for the first case, is at its null position for the second, and is stretched upward from its null position for the third.

In the first and third cases the deflective force $\mathbf{F}'$ was not measured, but from the photographs it is seen to be downward in the first case and upward in the third, corresponding to a positive and a negative submerged weight, respectively. In the second case, however, for which no deflection occurs, it is seen that $\mathbf{F}' = 0$, and we obtain from equation (35) the result that

$$\mathbf{F} = -m\mathbf{g}.$$

Then, since in this case the mass of the load was made equal to that of the displaced liquid, our final result is that

$$\mathbf{F} = -m\mathbf{g} = -m_l\mathbf{g}. \tag{36}$$

Thus, despite the fact that the water pressure has been completely excluded from the base of the loaded cylinder, we find experimentally that the force of buoyancy is still in precise accordance with the principle of Archimedes, and in complete agreement with equation (31) deduced from equation (1)

PLATE 1.—EXPERIMENTAL DEMONSTRATION OF BUOYANCY ON SOLID CYLINDER HAVING NO BASAL SURFACE POROSITY

FIGURE 1.—General view of apparatus
FIGURE 2.—Load greater than weight of displaced liquid

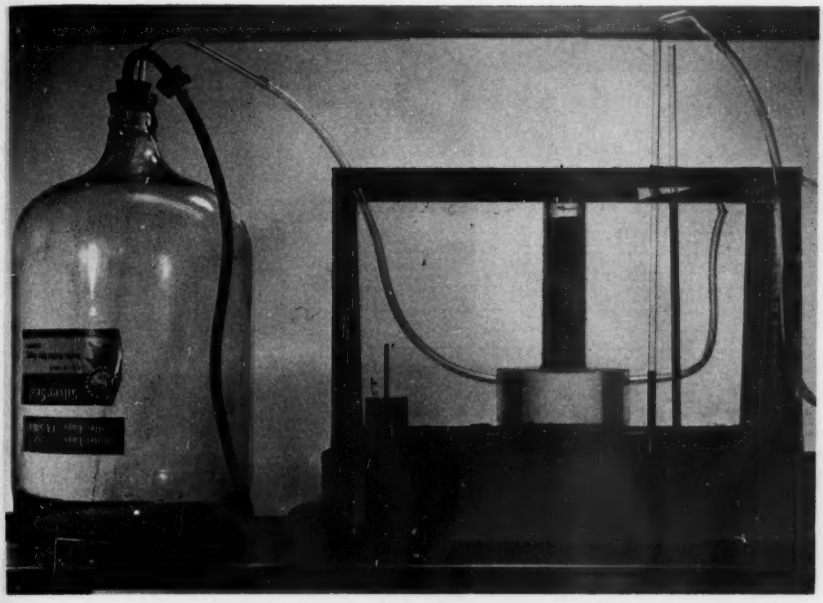

FIGURE 1

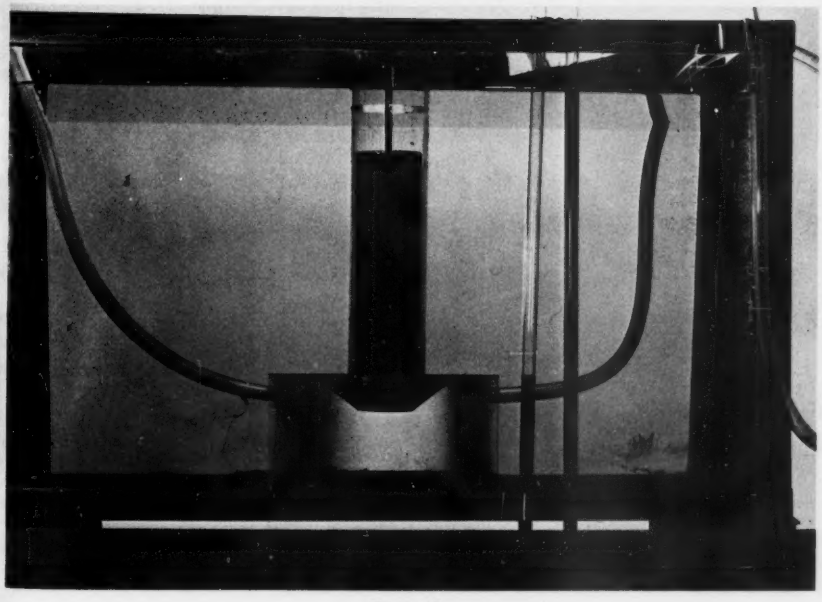

FIGURE 2

EXPERIMENTAL DEMONSTRATION OF BUOYANCY ON SOLID CYLINDER HAVING NO BASAL SURFACE POROSITY

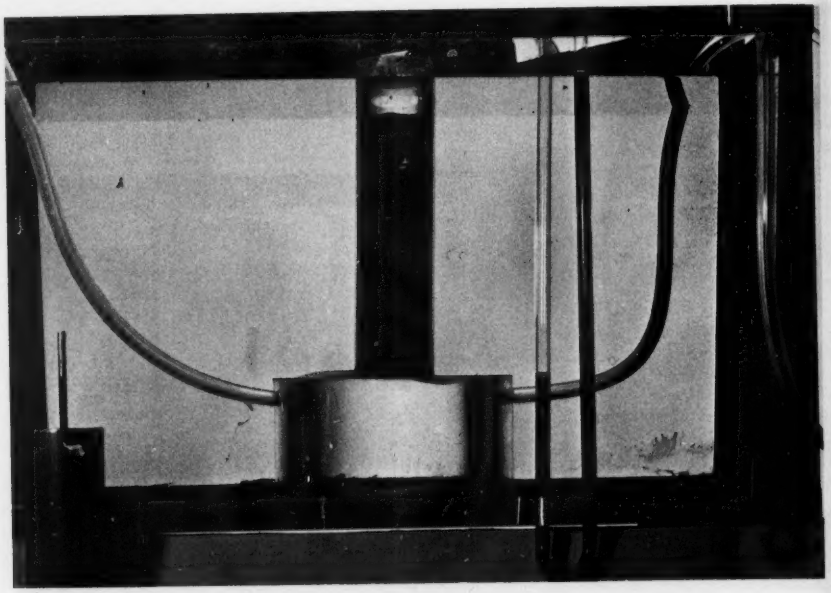

FIGURE 1

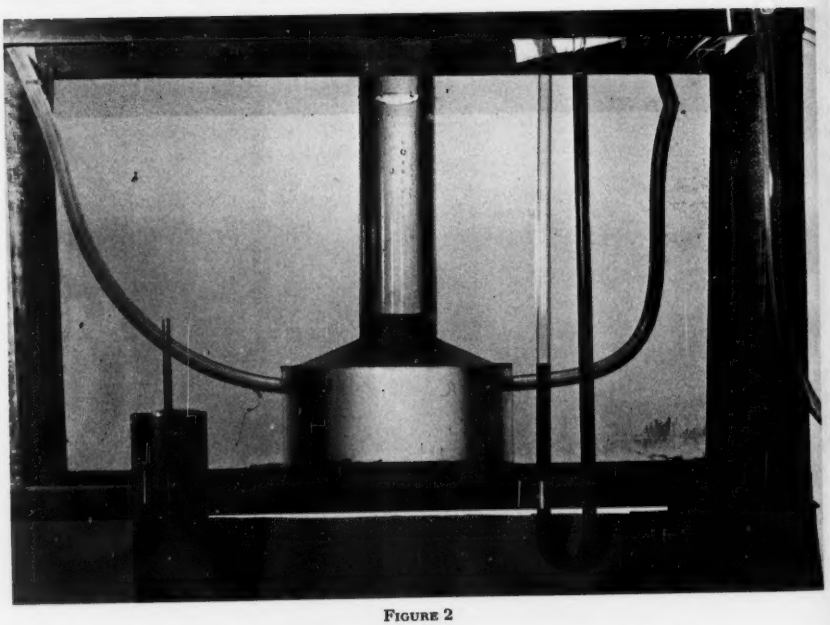

FIGURE 2

EXPERIMENTAL DEMONSTRATION OF BUOYANCY ON SOLID CYLINDER HAVING NO BASAL SURFACE POROSITY

herein; but it does not agree with the theory of Terzaghi.

Results Obtained From Triaxial Tests

The further question of whether the failure of porous rocks and soils in response to shear stresses is dependent solely upon the effective stress Ω, or whether the neutral stress Π is also involved, can best be determined empirically by experiment. This is most effectively done by the triaxial tests on jacketed specimens on the exterior of which the greatest and least total principal stresses S_1 and S_3 are applied, and in the interior of which the neutral stress Π is produced by means of a pore pressure p.

In this case, if we neglect the very small stresses due to the weight of the specimen and its pore fluid, the pressure p is constant inside the specimen, and grad p is zero. The neutral stress Π, therefore, exerts no directed forces which tend to separate the blocks of rock lying on opposite sides of a plane of potential shear failure. This is true because on each such block

$$\mathbf{F}_\Pi = -\iiint_V \text{grad } p \, dV = 0,$$

and the assumption that the fluid pressure p operating over an area $f_b A$ of the potential slip surface does exert such a force is another aspect of the fallacy of equation (2).

However, the amassed data of soil mechanics on loose soils, and the tests of Douglas McHenry (1948) on concrete, and of John Handin (1959) on solid rocks, as described by the authors on pages 140–142, do show that the failure of such materials in response to the effective stress is in good agreement with the Mohr-Coulomb law. Moreover, such failure is independent, within the limits of accuracy of the experiments, of the neutral stress Π.

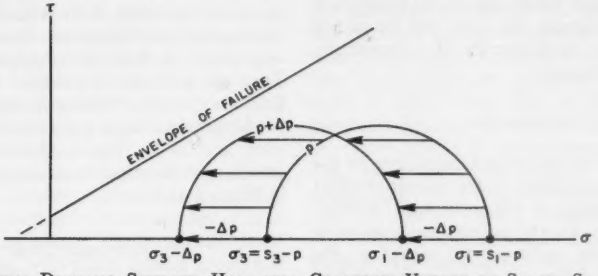

FIGURE 5.—MOHR DIAGRAM SHOWING HOW, FOR CONSTANT VALUES OF S_1 AND S_3, ROCK FAILURE CAN BE INDUCED BY INCREASE OF PORE PRESSURE

Accepting this empirical fact, the manner in which increase of the pore pressure can facilitate shear failure, even though the total stress remains constant, can perhaps more clearly be seen by reference to the Mohr diagram in Figure 5. Here, for a rock having a pore pressure p and the constant principal total stresses S_1 and S_3, a Mohr circle is shown for the principal effective stresses

$$\left.\begin{aligned} \sigma_1 &= S_1 - p, \\ \sigma_3 &= S_3 - p, \end{aligned}\right\} \tag{37}$$

which is short of the limit of failure defined by the Mohr envelopes.

Now, without changing S_1 and S_3, let the pore pressure be increased from p to $p + \Delta p$. The effective stresses will be changed to the new values

$$\left.\begin{aligned} \sigma_1 + \Delta\sigma_1 &= S_1 - (p + \Delta p), \\ \sigma_3 + \Delta\sigma_3 &= S_3 - (p + \Delta p). \end{aligned}\right\} \tag{38}$$

Then, when we subtract equations (37) from (38), it is seen that

---

PLATE 2.—EXPERIMENTAL DEMONSTRATION OF BUOYANCY ON SOLID CYLINDER HAVING NO BASAL SURFACE POROSITY

FIGURE 1.—Load equal to weight of displaced liquid
FIGURE 2.—Load less than weight of displaced liquid (cylinder empty)

$$\Delta\sigma_1 = \Delta\sigma_3 = -\Delta p. \tag{39}$$

Thus it is seen that, if the pore pressure is increased while the total stresses are kept constant, the Mohr circle for the effective stresses remains constant in diameter but is translated in the direction of decreasing compressive stress by a distance equal to the negative of the pressure increase. Obviously, if the pressure is increased sufficiently the circle will be moved until it becomes tangent to the Mohr envelopes, and failure will occur.

Conclusion

In the light of the critic's objections to the authors' earlier analysis of the buoyancy effect and other aspects of the mechanics of fluid-filled porous rocks, the authors have seriously re-examined the problem with regard to the validity both of the critic's arguments and of their own analysis and conclusions. As a result of this re-examination they find themselves in complete agreement with Terzaghi's resolution of the total stress field in a fluid-filled porous rock into a *neutral* and an *effective* component, where the neutral stress is equal at every point to the fluid pressure p as macroscopically determined. They do not agree, however, that the buoyancy effect of this neutral component is given by

$$\mathbf{F} = -\iint_A p f_b \, d\mathbf{A}, \tag{2}$$

or that it depends in any other manner upon the surface porosity.

While, in the authors' opinion, this equation is erroneous, its widespread use in soil mechanics and in other branches of civil engineering does not ordinarily lead to erroneous results because in practice it is found necessary to assign to f_b the value of unity in order to make the equation agree with experimental data. When this is done equation (2) reduces to the correct result,

$$\mathbf{F} = -\iint_A p \, d\mathbf{A}. \tag{1}$$

The authors merely wish to reiterate that, had the theory been valid in the first place, this expedient would not have become necessary.

Acknowledgments

In their re-examination of the problem under discussion, the authors have had the benefit of the advice, assistance, and criticism of several members of the scientific staff of the Exploration and Production Research Division laboratory of Shell Development Company. They are particularly indebted to Mr. Henry Rainbow for an extensive independent review of the problem from a mathematical point of view, and to Dr. Eugen Merten for the construction of the experimental apparatus shown in Plates 1 and 2.

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Shell Development Company, Exploration and Production Research Division, Houston, Texas; United States Geological Survey, Washington, D. C.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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STEREOSCOPIC-PAIR PROJECTION OF AERIAL PHOTOGRAPHS IN MAP COMPILATION

BY R. R. COATS

This note describes two methods for transferring information from sterographic pairs of aerial photographs to topographic base maps. Both methods require the use of a variable-ratio reflecting projector, of the Saltzman type, and both are useful only to persons capable of naked-eye stereoscopy.

The methods make the stereoscopic model appear nearly coincident with the topographic map, so that the lines on the map appear as if drawn on a relief model. The data may therefore be compiled with much greater speed and accuracy than by single-photo projection. The methods are most useful in areas of moderate relief; in areas of very low relief, with sufficient culture for control, they offer no advantages over single-photo projection, and their accuracy is diminished in areas of high relief, in proportion as the perspective view furnished by the photograph departs from an orthographic representation. Distortions produced by tip and tilt still remain.

In the first method, the two photos that form the stereoscopic pair are placed on a backing sheet in such a relation that four points, two on each photograph, fall on the same straight line. These points are the principal point of each photo and the transferred position of the principal point of the other photo. The separation of the two pictures, measured by the distance between the images of the corresponding points, is governed by the amount of scale reduction. For 1:1 compilation it should be, at most, slightly greater than the interpupillary distance of the operator. For a reduced compilation, the separation may be larger than the interpupillary distance in inverse proportion to the reduction ratio—*e.g.*, if the compilation is at half the picture scale, the pictures may be separated by twice the interpupillary distance. Each picture is held in place by a single strip of masking tape along the edge normal to the line connecting the principal points, and farthest from the area of overlap. This makes it easy to change the direction of shingling of the pictures, from left on top to right on top, or vice versa, so that all the area of overlap may be seen. The backing sheet with the photographs in position is placed in position in the projector, so that the projected position of the line connecting the principal points is parallel with the eye base of the operator. When the images of the two photographs are projected on a blank sheet of paper, they may be viewed stereoscopically just as the two photographs might be viewed directly to form a stereoscopic image. The model must be viewed obliquely, of course, and this obliquity creates the illusion that the hills are leaning toward the operator, but accuracy of compilation is not diminished thereby. Any annoyance created by the illusion may be minimized by reversing the map and photographs to bring the far side of the projected image nearest to the observer. When the observer has accustomed himself to viewing the projected images as a stereomodel, the paper should be moved out of the way, and the map should be shifted on the table to make it coincide as nearly as possible with the image of the photograph bearing the information to be transferred. The image of the other photograph continues to be projected on the blank paper, since it is not, of course, in register with the map. Generally some readjustment of position of the map or of the scale of projection, or of both, will be necessary to preserve the best possible fit as the transfer of information proceeds outward from the initial point. What the observer sees is the semblance of a relief model inscribed with the information printed on the base map. The method appears to be usable for compilations at ratios ranging from 1:1 to a 3:1 reduction; the scale of reduction is limited by the visibility on the projected image of the lines drawn on the photo.

This method of compilation, which may be termed stereoscopic-pair projection, seems as much superior to single-photograph projection as field mapping with stereoscopic pairs is to field mapping with single photographs. It increases accuracy of transfer as much as field use of stereoscopic pairs increases accuracy of location. The increased accuracy results from the increased amount of control over the position of the projected image in relation to the map that results from the existence of the stereoscopic image.

The second method of compilation is a

special case of the first. It suffers from the severe limitation that it is usable only for compilation at a ratio of nearly 1:1; its sole advantage is that the stereoscopic image is somewhat brighter and clearer than in the first method. The procedure is as follows: The photograph with the material to be compiled is placed in the object holder of the projector with the flight line in such a position that it will, on projection, be parallel to the eye base of the operator. The topographic map is placed on the plotting table in approximately the correct position. That part of the area of the projected photograph which would be hidden, in ordinary naked-eye stereoscopy, by the overlap of the other member of the stereopair, is covered with white paper. The second picture is then placed on the plotting table in such a position that the image of the blank paper falls on it. The projected photo and the photo on the plotting table may be viewed together as a stereomodel by the usual method of naked-eye steroscopy. Since the photograph containing the information to be compiled must be projected to the topographic map, a reversal of the direction of overlap, used in normal naked-eye stereoscopy, is not possible. What we have is a stereomodel like that which would be produced if the photograph on the plotting table were to overlap directly the part of the projected photo that is concealed by white paper. If we were to attempt to reverse the direction of overlap, the white paper would have to be used to mask part of the photograph on the plotting table, and the information on the projected photograph would be compiled on this paper, instead of on the map. All information to be compiled must therefore be plotted on a strip on the central part of each photograph. The width along the flight direction of the strip thus usable will vary with the interpupillary distance of the observer, but 3-inch widths seem fairly easy to attain, even when the interpupillary distance is slightly less than that.

Either method may also be used as a quick although crude method of catching gross errors in a topographic map. Errors of the map stand out vividly and can be detected more certainly as well as more quickly, when stereomodel and map approximately coincide, than when the eye must travel back and forth from model to map.

U. S. Geological Survey, Menlo Park, Calif.
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Publication Authorized by the Director, U. S. Geological Survey

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PHOSPHATIZED WOOD FROM THE PACIFIC SEA FLOOR

BY EDWARD D. GOLDBERG AND ROBERT H. PARKER

A recent expedition of the Scripps Institution of Oceanography, Tuna Oceanographic Cruise TO-58-2, Station D-7, dredged a sample of phosphatized wood from a 410-meter terrace in the Gulf of Tehuantepec (Dredge initiated at Lat. 14° 25′ N. and Long. 93° 05′ W. and concluded at Lat. 14° 27′ N. and Long. 93° 11′ W.). This wood has been identified as a dicotyledon (Elso S. Barghoorn, personal communication). A carbon-14 age determination made by Professor Hans Suess of this Institution on the organic wood phases indicates an age of greater than 28,000 years.

An X-ray diffraction spectrum of the mineralized portion revealed the phosphatic material to be similar to marine carbonate-fluorapatites, $Ca_{10}(PO_4, CO_3)_6F_{2\text{-}3}$, in which excess positive charge resulting from substitution of carbonate for phosphate is balanced by excess fluorine or hydroxyl groups. The mineral is a hexagonal network composed of phosphate tetrahedra, with the anions fluorine and hydroxide, and the divalent cation calcium within the structure (Altschuler, Clarke, and Young, 1958). The ratio of the uranium and thorium contents in the fossilized wood, 13 and 0.2 ppm respectively, is in agreement with the values reported by the above authors for marine apatites and is of concern here inasmuch as certain aspects of the marine geochemistry of these elements can be interpreted through studies of their occurrence in phosphatized wood and other calcium phosphates.

The orientation of the wood on the sea bottom is established by clay that adheres to one part of the specimen, the buried portion, but not to the remainder, the portion in contact with sea water. The phosphatization occurred only in that portion of the wood which was above the sediment–water interface; that part of the wood buried in the deposit was essentially free of phosphate infiltration.

The phosphate, as well as the other constituents of the apatite, were clearly derived from the overlying sea water, which at the 400-meter level in this area contains maximal values of phosphorus, of the order of 3 micromoles per liter (Fig. 1). The phosphate content of woods is only about 0.02 grams of P_2O_5 per 100 grams of dry weight material (Specter, 1956). For every molecule of oxygen in the sea water there is a molecule of phosphate and a few thousand molecules of calcium.

This area is notably productive with respect to plant life. The high fertility is reflected in the low values of oxygen below the photosynthetic zone where this gas has been consumed in the combustion of organic matter arising from plant and animal remains. The overlying waters have an exceedingly low oxygen content (between 0 and 0.1 ml. of O_2/liter) or an upper value of about 5 micromoles per liter (Fig. 1). The low oxygen values have a two-fold significance. First, these low tensions probably result in an incomplete combustion of the organic matter to forms less oxidized than CO_2. Hence, only minor contributions to that carbonate content of the sea water can be expected from the oxidation. Second, the absence of the apatite from the buried wood indicates that the combustion did not take place throughout the entire sample. The oxygen was apparently completely used up in the more easily penetratable, exposed portions.

The conditions of phosphatization can be reconstructed from the above evidences and previous investigations of the occurrences and properties of marine apatites. By the time the wood reached the deposit site, it was completely penetrated by sea water. Dissolved oxygen of the sea water reacted with the more readily oxidizable organic components of the wood with a subsequent reduction in Eh. This also leads to a decrease in pH, and this in turn diminishes the carbonate content. Although the hydrogen ion concentrations controlling the precipitation of calcium carbonate and calcium phosphate are remarkably similar, the high phosphate content of the water apparently turns the tide in favor of apatite deposition. Such an event has been postulated by Krumbein and Garrels (1952) who point out that phosphorite may well form in restricted basins in which the pH is relatively low—*i.e.*, deposits near or at anaerobism and with pH slightly lower than normal sea-water values of around 8.

The high uranium/thorium ratio, around 65, provides further evidence for delimiting the physico-chemical conditions of the environ-

ment. Altschuler *et al.* (1958) have demonstrated that the phosphatic nodules from the Pacific Ocean contain predominantly tetravalent uranium which is presumed to substitute directly for calcium in the apatite lattice; the

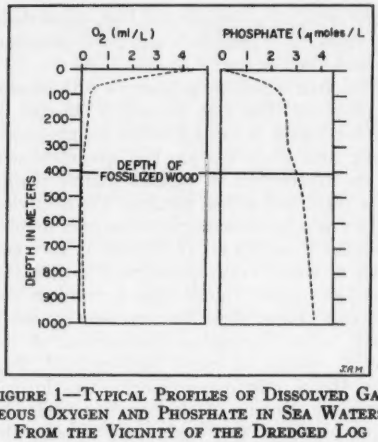

FIGURE 1—TYPICAL PROFILES OF DISSOLVED GASEOUS OXYGEN AND PHOSPHATE IN SEA WATERS FROM THE VICINITY OF THE DREDGED LOG
Data from Tuna Oceanographic Cruise TO-58-2

ionic radius of U^{+4} is 0.97 Å, nearly identical to that of Ca^{+2} of 0.99 Å. Uranium enters the ocean in the hexavalent state, probably complexed as the carbonate (Starik and Kolyadin, 1957). The reduction of uranium would be enhanced in those phases of the deposit out of contact with the aerated water and approaching or at anaerobism. The simultaneous reduction of uranium with the formation of the apatite structure which easily accommodates the tetravalent form explains the enrichment of uranium in these nodules. The ionic radius of tetravalent thorium of 1.02 Å explains the capture of this ion from sea water by the apatite structure, in a way similar to that of tetravalent uranium. Whereas the *total* concentration of uranium in sea water is about 3 micrograms per liter, however, thorium concentrations have been delimited to this date only by a maximum value of 10^{-2} micrograms/liter. One can thus conclude that the activities of Th^{+4} and U^{+4} in the reducing environment of apatite formation are of the same order of magnitude as the concentrations of these elements in sea water.

These views are not in conflict with present opinion on the deposition of phosphate in the marine environment (McKelvey *et al.*, 1953). Marine apatites today are found in eastern oceanic coastal deposits where upwelled waters give rise to high productivity of organic matter. The rapid accumulation of this organic matter on the sea floor can produce an environment similar to that of our wood sample.

This work was supported by the Office of Naval Research under contracts with the University of California.

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SCRIPPS INSTITUTION OF OCEANOGRAPHY, UNIVERSITY OF CALIFORNIA, LA JOLLA, CALIFORNIA
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, DECEMBER 1, 1959

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LOAD-CAST TERMINOLOGY AND ORIGIN OF CONVOLUTE BEDDING: SOME COMMENTS

BY C. H. HOLLAND

H. H. Sullwold, Jr. (1959) objects to the use of the term "load cast" and proposes to replace it by three new terms: "load pocket", "load wave", and "load fold". His first objection to the original, and well-established, "load cast" is that "the cast is a third-order term". Does this imply that the best terms are "first-order"; is, for example, "sandstone" a poorer term for being "second-order", implying "sand" and later (by lithification) "stone".

His second objection to "load cast" is that it seems to apply generally both to the bulges of sand projecting downward into the underlying finer-grained sediment *and* to the plumes of mud which penetrate upward into the sand. Sullwold evidently believes, however, that the meaning of the term is weighted in favor of the first of these features, for the root of his objection is that it is the second feature which is the more easily seen in sections of poorly consolidated strata, such as the Tertiary rocks of California from which his illustrations are drawn. He proposes to employ the term "load pocket" for the bulges of sand and "load wave" for the plumes of mud.

The present writer does not believe that "load cast" is generally taken to apply to both these features. Surely it is usually applied to the sand bulges, and a satisfactory term, "flame-structure", is already available for the mud plumes. These terms have been clearly defined, for example, by Kelling and Walton (1957), and there seems to be no good reason for their replacement.

Sullwold's reference, in his last paragraph, to the origin of convolute bedding also calls for comment. It is true that if by submarine erosion of a sequence such as the one shown in his Figure 3 of Plate 1, all traces of the overlying coarser sediment and its downward projections were removed, then the remaining "load folds", covered by subsequent deposition, might resemble slightly developed convolute bedding. But this is very far from saying that the folds might *be* convolute bedding (Sullwold, p. 1248). A diagnostic feature of convolute bedding, which has been described for example by F. P. H. W. Kopstein (1954, Thesis, Univ. Groningen), Ten Haaf (1956), and Holland (1959), is that the convolutions die out not only downward but *frequently upward as well*. Whatever may be the origin of convolute bedding it is not possible that "load folds" of the kind described could ever die out upward in this way. The "load folds" develop upward into flame structures or are covered by the overlying coarser sediment. If all traces of the coarser sediment (and hence also of the flame structures) were removed by submarine erosion the underlying "load folds" would inevitably be truncated. Convolute bedding is not formed in underlying sediment by a turbidite descending upon it; rather it is formed *within* the turbidite itself, at the time of deposition of the latter.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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LOAD-CAST TERMINOLOGY AND ORIGIN OF CONVOLUTE BEDDING: FURTHER COMMENTS

BY HAROLD H. SULLWOLD, JR.

The preceding comments by C. H. Holland with reference to my earlier Note (Sullwold, 1959) are appreciated but call for comment.

(1) I did not advocate replacing "load cast" but introduced three additional terms of equal status. I actually admire the term load cast but pointed out that it was third order in the sense that the cast was preceded by the depression (mold) and the load. No stand is intended on which order is "best".

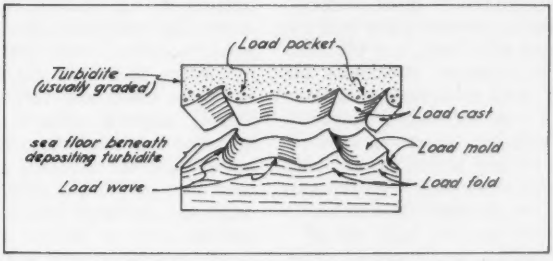

FIGURE 1.—DIAGRAM ILLUSTRATING TERMS OF LOAD DEFORMATION

(2) Cast means that which is formed in a mold and also means throw and therefore *seems* to apply to both the nether bulges and the plumes (load waves or flame structures) which appear to have been thrown up. I fully realize that load cast is a cast of a depression in the mud of the sea floor caused by a load and should be so used.

(3) Unfortunately I inexcusably overlooked the term "flame structure" in the literature. This term, incidentally, was not proposed by Kelling and Walton (1957) but by Walton (1956). Load wave and flame structure are not necessarily synonymous, as the former refers to smooth upward bulges as well as to tenuous breaking wave or flame shapes.

(4) Load pocket is not meant to replace load cast but to supplement it. I conceive of load casting as the configuration of the undersurface of the turbidite (seen also as a wavy line in cross section), whereas the load pocket is the material within the load cast (seen only in cross section). These are two distinct concepts each requiring identity. Other concepts include the load mold, load fold, and load wave as shown in Figure 1. A reasonably consistent descriptive classification of five genetically related features results. Ph. H. Kuenen (personal communication, November 7, 1959), in referring to the Note under discussion, states "... I can agree with your suggestions to add terms in cases such as you describe."

(5) Holland's comments on convolute bedding are well taken. I agree that convolutions which die out upward could not be load folds. However, I have rarely seen them die out upward, and both illustrations of the feature in the paper which defined the term (Kuenen, 1953) appear to show erosion at the top. This same paper implies that the mode of origin is not known with certainty. If the term convolute bedding is descriptive, referring to internal contortions which may be distinguished from slump structures, but whose origin is not certain, then some bevelled load folds may *be* convolute bedding. The writer prefers this concept. If convolute bedding is a genetic term whose use implies knowledge of origin, then we must be very cautious in using it, and a new descriptive term is needed. As a matter of fact, Kuenen included in his definition that convolute bedding is "... produced during the deposition in the highly mobile sediment." I would hope that he would not object to using the term in the descriptive sense. In the previously cited letter to the writer Kuenen pointed out the particular circumstances of erosion that would be required to

produce confusion between load folds and convolute bedding, tacitly admitting the possibility but implying scarcity of occurrence. I offered the suggestion and still believe that bevelled load folds might be mistaken for bevelled convolute bedding. Surely this is a possibility and perhaps a common one.

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GEOLOGIC TIME SCALE

BY HENRY FAUL

ABSTRACT

Evidence is gradually accumulating that the presently accepted (Holmes-Marble) geologic time scale should be lengthened, and that this extension should be greatest in the Paleozoic era. Age determinations on glauconites, however, contradict this evidence and generally support the present scale. A review of available data indicates many uncertainties. It still seems doubtful to attempt the construction of a new time scale until the key points can be established firmly enough to be useful.

Numerical values of time can be attached to points in the stratigraphic time scale whenever a rock satisfies three rather stringent requirements: (1) the rock must contain at least one mineral that is a closed system to both parent and daughter isotopes; (2) the mineral must have crystallized at a time that can be geologically related to the deposition of fossiliferous sediments, and (3) the rock must not have been subjected to subsequent metamorphism severe enough to alter parent/daughter ratios. The first requirement is, of course, the basic premise tacitly assumed whenever one calculates an age by radioactive decay.

Geochronologically Useful Minerals

Zircon, monazite, xenotime, uraninite, thorianite, coffinite, biotite, muscovite, lepidolite, K feldspar, and glauconite all have been used in age determination. Tilton and his coworkers (1957) have shown that zircon often gives concordant ages by the three uranium-lead methods. The thorium-lead age is usually low, and the reason for this discrepancy remains obscure. Uraninite also often gives concordant ages, but the rarity of this mineral curtails its usefulness. On the other hand, zircon and the uranium minerals may give discordant ages by the uranium and lead methods. The usual pattern is $U^{238}/Pb^{206} < U^{235}/Pb^{207} < Pb^{207}/Pb^{206}$. There are ways of adjusting such ages into concordance (Wickman, 1942; Stieff and Stern, 1956; Wetherill, 1956a; 1956b), and unique solutions are obtained for some groups of data (Wetherill, 1956a; 1956b; Gerling, 1958); but as a rule the calculations have little meaning for rocks younger than about 500 million years, and that includes all rocks related to fossiliferous sediments. One way of explaining the difficulty is to point out that the Concordia curve (Wetherill, 1956a; 1956b) does not have much curvature over the span of less than 500 m.y., and most analytical data are not sufficiently accurate to permit precise numerical calculations here. The original-lead correction is often the root of the difficulty. We may generalize that discordant lead ages provide some insight into the past history of a rock but are of little use in the establishment of a time scale for Paleozoic and younger rocks.

In this light we must consider the results of the Larsen (Pb/α) method on zircon (Jaffe *et al.*, 1959; Gottfried *et al.*, 1959). The method offers simplicity and speed, but no means of recognizing a discordant age. The Pb/α ratio reflects primarily the Pb^{206}/U^{238} ratio in normal zircons so that, if the zircon is concordant, the Pb/α age will be very nearly correct; when the zircon is discordant, however, the Pb/α age may be seriously in error.

Aldrich *et al.* (1958) have shown that muscovite and biotite usually give concordant ages by the K/Ar and Rb/Sr methods. Wherever the K/Ar and Rb/Sr ages clearly disagree, most workers prefer to accept the Rb/Sr ages as more nearly correct. In some pegmatites (Wasserburg *et al.*, 1959; Polkanov and Gerling, 1958) the Rb/Sr age of the muscovite is greater than other measured ages, for unknown reasons, but in granites the muscovite and biotite Rb/Sr ages generally agree.

K/Ar measurements on old feldspar give ages that are almost always too low, for feldspar does not retain all its radiogenic argon. The argon loss is variable and may be as high as 40 per cent (Wetherill *et al.*, 1955; Wasserburg *et al.*, 1956; Goldich *et al.*, 1957). Age measurements on feldspathic whole rocks, therefore, can be expected to give erroneously low values if much of the potassium in the rock is present in feldspar (Gentner and Kley, 1958; Mehnert, 1958).

Rb/Sr measurements can serve as a basis for accurate ages only when the ratio of radiogenic strontium^{87} to common strontium^{87} in the mineral is large enough to permit a meaningful calculation. In practice that means a value of at least 0.1. Confidence is greater when the ratio is larger. Many feldspars and glauconites do not meet this requirement. To a certain degree, a favorable radiogenic/common Sr ratio is a matter of clean mineral separation, but even with the greatest care it is often impossible to calculate a meaningful Rb/Sr age.

Glauconite, like feldspar, also loses radiogenic argon in varying proportion, largely as a function of temperature (Evernden *et al.*, 1959; Amirkhanov *et al.*, 1957). Consequently, K/Ar ages measured on glauconite can be expected to be lower than the real age of the rock.

Stratigraphic Correlation

Most of the geochronologically useful minerals occur in igneous rocks, and it is usually difficult to establish sufficiently narrow stratigraphic limits of the time when an igneous rock crystallized. Good exposures are necessary to reveal the contacts, but stratigraphic correlation is usually complicated by contact metamorphism which tends to obscure the fossils and stratigraphic relations in the rock that determines the lower stratigraphic limit. The upper contact also must be seen well enough to prove that it is unconformable.

Plutons without overlying sediments nevertheless can be useful in giving a minimum age for sediments they intrude. Tectonic evidence may indicate that intrusion probably followed soon after deposition of the sediments that now make up the hornfels rim. This is the case in Jackman, Maine (Hurley, Boucot, *et al.*, 1959), Katahdin, Maine (Boucot, 1954), and in the Oslo region (Faul *et al.*, 1959), where age measurements give minimum values compatible with other measurements on more closely bracketed plutons.

The stratigraphic age of most pegmatites is particularly difficult to determine. Of the five points used by Holmes (1947) to establish his famous time scale, two are pegmatites with excellent concordant lead age measurements (Nier *et al.*, 1941) but with only a nebulous stratigraphic correlation. The Pala point listed below is suspect for similar reasons.

Some volcanic ash falls contain biotite and K feldspar. The stratigraphic position of bentonites is rarely to be questioned, and it is petrologically reasonable to assume that the biotite in these rocks crystallized (or was hot enough to lose its argon) shortly before the ejection and deposition of the material. The effect of this process on the Rb/Sr ratio in mica or feldspar is incompletely known.

Glauconite is fairly common in sedimentary rocks and would be extremely useful if it were not for the problems of the time of its crystallization, argon loss, and excessive common strontium contamination, discussed above.

Metamorphism

Finally, one must consider all possibilities of metamorphism which could have caused even partial recrystallization of the minerals. The effects of metamorphism on different age ratios and different minerals are widely different. Zircon can survive fairly drastic metamorphism, and feldspar may retain its Rb/Sr ratio under similar conditions (Tilton *et al.*, 1958). Argon is lost fairly quickly at elevated temperatures (Reynolds, 1957; Evernden *et al.*, 1959). The resistance of micas to metamorphic processes is not well known. The pegmatite ages already mentioned (Wasserburg *et al.*, 1959) may indicate that muscovite retains its radiogenic strontium even when biotite loses it and when all micas lose argon.

Points in the Time Scale

In the light of these considerations, we may now review the available age determinations on rocks that fulfill at least some of the stated requirements, recomputed, where necessary, with the decay constants reported by Aldrich and Wetherill (1958; Aldrich *et al.*, 1958). For the Rb half-life we use the new value of 4.7 $\times 10^{10}$ years (K. F. Flynn and L. E. Glendenin, preliminary report, 1959).

Middle or late EOCENE

Carbonatite with large (10 cm) biotite crystals occurs in the composite Rocky Boy stock in western Hill County, Bearpaw Mountains, Montana. The stock is correlated with intrusives that cut sediments of Wasatch (early Eocene) age. Associated volcanic rocks are interbedded with sedimentary layers bearing fossil plants of middle and late Eocene age (Pecora, 1957).

K/Ar (Rocky Boy prospect biotite, Faul, Table 2) 52 ± 2 m.y.

LARAMIDE Mineralization (Post-Late Cretaceous)

This is one of the five tie points of the time scale of Holmes (1947, Gilpin County, Colorado). The ages are U^{238}/Pb^{206} on uraninite from veins near Central City, Colorado, that cut bostonite dikes petrologically correlated with other dikes that in turn intrude the Pierre shale of Late Cretaceous

age near the southwestern end of the porphyry belt. The complex relationship is summarized by Lovering and Goddard (1938, 1950, p. 44–47).

U^{238}/Pb^{206}
(Nier *et al.*, 1941, average of 2 samples) 58 m.y.
(U. S. Geological Survey, *see* Faul, 1954, p. 263, average of 4 samples) 63 ± 6 m.y.
(Wood mine, Kulp, 1955) 56 ± 5 m.y.

Probably Early LATE CRETACEOUS

The lepidolite-bearing pegmatite at Pala (40 miles north of San Diego, California) lies in the San Marcos gabbro which is part of the Southern California batholith. By extrapolation to rocks in Baja California, the batholith is tentatively dated as Early Late Cretaceous (Larsen *et al.*, 1958, p. 48).

Rb/Sr
(lepidolite, Webster *et al.*, 1957, recomputed) 96 m.y.
(lepidolite, Aldrich *et al.*, 1958, recomputed) 100 m.y.
(lepidolite, Herzog *et al.*, 1956, recomputed) 114 m.y.
K/Ar
(lepidolite Aldrich *et al.*, 1958b) 89 m.y.
(lepidolite, Reynolds, 1957) 87 m.y.
(K feldspar, same) 88 m.y.
(lepidolite, Baadsgaard *et al.*, 1957) 96 m.y.

Early LATE CRETACEOUS (Post-Albian, pre-Maestrichtian)

In northwestern Baja California, intrusive rocks of mostly dioritic composition invaded and metamorphosed fossiliferous limestones and volcanic sediments of Albian age. A fossiliferous sedimentary sequence of Maestrichtian age nonconformably overlies the plutons as well as the rocks they have metamorphosed (Silver *et al.*, 1956, and preliminary report, 1956).

U/Pb (La Grulla granodiorite, concordant age on monazite, Silver *et al.*, preliminary report) 115 m.y.

Probably post-JURASSIC and pre-EARLY CRETACEOUS (Valanginian)

The Shasta Bally batholith in the Klamath Mountains, California, intrudes Mississippian sedimentary rocks. Structural evidence suggests that it was emplaced in early Late Jurassic time, or later. It is overlain by essentially untransported debris correlated with fossiliferous Valanginian (Early Cretaceous) sedimentary rocks (Curtis *et al.*, 1958; F. G. Wells private note).

K/Ar (biotite, chlorite present, Curtis *et al.*, 1958) 127 m.y.

LATE JURASSIC (Oxfordian-Kimmeridgian) or later

In central California, the Guadalupe Mountain quartz monzonite intrudes the Mariposa formation (Oxfordian-Kimmeridgian) and older volcanic rocks. About 100 miles northwestward, the Horseshoe Bar quartz diorite is surrounded by the Rocklin granodiorite, which probably also intrudes rocks correlated with the Mariposa (Curtis *et al.*, 1958).

K/Ar
(biotite, Guadalupe Mtn.) 136 m.y.
(biotite, Horseshoe Bar) 136 m.y.
(biotite, Rocklin) 125 m.y.
(muscovite, Rocklin) 124 m.y.

EARLY PERMIAN or later

Granite and nordmarkite intrude the Oslo rhomb-porphyry lavas which are interbedded with fossiliferous lake sediments of early Permian age that are correlative with the German *Rotliegendes*.

K/Ar
(biotite from the Drammen granite, Faul *et al.*, 1959) 259 m.y.
U^{238}/Pb^{206}
(zircon from the Oslo nordmarkite, Faul *et al.*, 1959) 260 m.y.

LATE PENNSYLVANIAN or later

Pennsylvanian rocks of the Narragansett Basin, Rhode Island, are intruded and metamorphosed. The metamorphism may have followed soon after the end of deposition (Hurley *et al.*, in press).

Rb/Sr
(biotite from granite, recomputed) 244 ± 13 m.y.
K/Ar
(biotite from granite, average of 2 samples) 234 ± 11 m.y.
(biotite from schist) 250 ± 12 m.y.
(whole slate, average of 3 samples) 248 ± 13 m.y.

LATE CARBONIFEROUS (post-Westphalian) or later

The Dartmoor and Land's End granites in Cornwall intrude Lower Westphalian (Lower Pennsylvanian) sedimentary rocks (Brammall, 1926), but their upper limit is uncertain. "No rocks occurring in the [Permian] breccias can be ascribed with certainty to the Dartmoor granite . . . ." (Dewey, 1948, p. 40), but detrital minerals probably of Dartmoor origin are identified in Weald (Lower Jurassic) sediments (Groves, 1931).

Dartmoor:
K/Ar
(biotite from granite; Mayne *et al.*, 1959, recomputed) 295 m.y.
(biotite sample furnished by K. I. Mayne; Faul, Table 2) 290 m.y.

Land's End:
K/Ar
(biotite from granite, Mayne *et al.*, 1959, recomputed) 330 m.y.

EARLY CARBONIFEROUS (Dinantian, pre-Visean)

In northeastern France, the Vosges granites intruded and metamorphosed Tournaisian sedimentary rocks. Overlying fossiliferous rocks of early Visean age were not affected (Jung, 1928, p. 34).

Rb/Sr (biotite, average of 3 samples; Carnegie Institution, 1957, recomputed) 322 m.y.
K/Ar (biotite, average of 4 samples; Faul, Table 2) 315 m.y.

LATE DEVONIAN

A biotite-bearing layer of bentonite, about 1 inch thick, is found in the upper part of the Dowelltown member of the Chattanooga shale of Late Devonian age in Tennessee. The layer is recognized as clearly volcanic in origin (Hass, 1948).

Rb/Sr (biotite, Adams *et al.*, 1958, recomputed) 385 ± 40 m.y.
K/Ar (biotite, Faul and Thomas, 1959) 340 ± 7 m.y.

MIDDLE DEVONIAN or later (post-Oriskany)

The Hog Island quartz monzonite near Jackman, Maine, intruded and contact-metamorphosed fossiliferous sedimentary rocks of Oriskany age to a cordierite hornfels (Hurley, Boucot, *et al.*, 1959).

Rb/Sr(biotite, recomputed) 362 m.y.
K/Ar
(biotite, average of 4 samples) 360 m.y.
(whole slate and hornfels, average of 4 samples) 365 m.y

EARLY or MIDDLE DEVONIAN

Late Silurian (middle Ludlow to lower Gedinne) fossiliferous slates near Calais, Maine, are intruded and metamorphosed. The intruding granites were unroofed and eroded, and one of them is now unconformably overlain by sandstone of Late Devonian age (D. H. Amos, 1959, Ph.D. thesis, Univ. Ill.; A. J. Boucot, private letter, 1959).

K/Ar (biotite, average of 4 samples, Faul, 1959, and Table 2) 404 ± 8 m.y.

LATE SILURIAN or later

The Cairnsmore of Fleet granodiorite (Kirkcudbrightshire, Scotland) intrudes Llandovery sediments, and the related Criffell-Dalbeattie pluton intrudes both Llandovery and Wenlock graywackes and shales (Phillips, 1956). Both plutons were unroofed in Late Devonian or Early Carboniferous time, judging by fragments in the sedimentary rocks of that age.

K/Ar (biotite, Cairnsmore, Mayne *et al.*, 1959, recalculated) (460) m.y.

MIDDLE ORDOVICIAN

A layer of volcanic ash occurs near the top of the Carters limestone, Stones River group, in the Middle Ordovician of Albama. It is exposed in mines and borings, and there it contains fresh euhedral biotite, sanidine, and some zircon (Faul and Thomas, 1959).

Rb/Sr
(biotite, Adams *et al.*, 1958, recomputed) 466 ± 50 m.y.
(biotite, Wenonah No. 8 mine, Tilton, private note) 447 ± 40 m.y.
K/Ar
(biotite, average of 2 samples, Pyne mine and Wenonah No. 8, Faul and Thomas, 1959) 419 ± 5 m.y.
(same, Curtis and Evernden, private note) 421 ± 5 m.y.
U^{238}/Pb^{206} (zircon, Tilton, private note) 445 ± 10 m.y.
U^{235}/Pb^{207} (Tilton, private note) (455 ± 10 m.y.)
Pb^{207}/Pb^{206} (Tilton, private note) (545 ± 45 m.y.)

Another similar tuff is found in the Eggleston limestone, Middle Ordovician, of Tennessee.

Rb/Sr (biotite, average of 2 samples, Adams *et al.*, 1958, recomputed) 437 ± 50 m.y.

The same ash bed extends far to the north. Zircon concentrates were prepared from bentonite samples taken at Alexandria, Dowelltown, and Nashville, Tennessee, and analyzed by Edwards and coworkers (1959) by the isotopic lead-uranium method.

U^{238}/Pb^{206}
(zircon, Alexandria) 452 m.y.
(zircon, Dowelltown) 446 m.y.
(zircon, Nashville) 438 m.y.

A sample of another bentonite was collected near Vonore (southwest of Knoxville), Tennessee, from the Middle Ordovician Bays formation (Edwards *et al.*, 1959).

U^{238}/Pb^{206} (zircon, Edwards *et al.*, 1959) 453 m.y.

LATE CAMBRIAN

Glauconite occurs in the Franconia sandstone of Late Cambrian age at Readstown and Sparta, Wisconsin, and in Goodhue County, Minnesota.

Rb/Sr (Readstown, Wisconsin, Herzog *et al.*, 1958, recomputed) 414 m.y.
K/Ar (Sparta, Wisconsin, Wasserburg *et al.*, 1956) 442 m.y.
(Goodhue County, Minnesota, Goldich *et al.*, 1959, average of 2 samples) 439 m.y.

Discussion

Some of these highly selected numbers satisfy our stated conditions reasonably well, but most of them fall short on one count or more.

Of the five points on which Holmes based his time scale, only one (Laramide) can be included now. The stratigraphically unimpeachable "Swedish Kolm" from the alum shale does not present a closed system, and all attempts to establish an age for it have failed. The stratigraphic limits on Holmes' remaining three points are too vague to make them useful.

Some of the other points quoted here are based on determinations by just one method in one laboratory. On biotites, one would naturally prefer to have both the Rb/Sr and the K/Ar ages, be they concordant or not. K feldspar also should be helpful where the radiogenic/common Sr ratio is favorable. If the rock contains sufficient zircon, then isotopic U/Pb analysis is a necessity, for it will either confirm the mica ages or give a glimpse of the past metamorphic history of the rock.

Reasonably complete analyses are available for only four of the new tie points (Jackman, Vosges, Alabama-Tennessee bentonite, and Franconia glauconite). Of these points, the Jackman age is only a minimum, and the Franconia age suffers from all the uncertainties inherent in glauconite ages.

The discordance in the ages of the Alabama tuffs is small but probably significant. The low argon ages could be explained by partial loss of argon and/or addition of potassium as a result

of slight early alteration of the mica. The discordance in the U^{238}/Pb^{206}–Pb^{207}/Pb^{206} ages may indicate the presence of some relic zircon from older rocks. The true age of the zircon may be obtainable from a systematic mathematical analysis of all the zircon data. For the moment one can say only that the Alabama bentonite was deposited not later than 420 million years ago and probably not earlier than about 450 million years ago.

The reported K/Ar age of the Cairnsmore of Fleet pluton is too great to fit even the broad scheme. Only additional analyses can clarify this apparent discrepancy.

In the Calais (Maine) area the granite body intruding the Silurian slates is petrographically slightly different from the body overlain by the Devonian sandstone. Contacts between the two granites are not exposed, but the K/Ar age of most of the granites in the area is the same, and we assume that the different granitic rocks are merely facies of the same igneous body. There is some evidence of possible metamorphism to the west, however, which may complicate the picture.

The upper stratigraphic limits of the Dartmoor and Land's End granites are still uncertain, in spite of extensive geologic work in that area.

The field relation of the Horseshoe Bar and Rocklin plutons is not clear. However, the measured age of the Horseshoe Bar sample supports the value for the Guadalupe Mountain quartz monzonite which clearly intrudes the Mariposa formation. The stratigraphic limits on the Pala pegmatite are uncertain, and the point is of limited use in the time scale.

The K/A–Rb/Sr ages of the glauconite from

TABLE 1.—HOLMES-MARBLE TIME SCALE AND SUMMARY OF NEW RESULTS

All in millions of years. Bold type indicates points based on better-than-average evidence, both geologic and experimental. Minimum ages are marked >.

Holmes-Marble time scale		NEW RESULTS	
	Tertiary	50	BEARPAW MTNS.
60		60	GILPIN CO.
	Cretaceous	95	PALA
130		115	LA GRULLA
	Jurassic	130	SHASTA BALLY
155		135	GUADALUPE
	Triassic		
185			
	Permian	>260	OSLO
210		>250	NARRAGANSETT
	Pennsylvanian	>290	DARTMOOR
235			
	Missisippian	**320**	VOSGES
265		340-385	TENN.
	Devonian	**>360**	JACKMAN, ME.
320		>400	CALAIS, ME.
	Silurian		
360			
	Ordovician	**420-450**	ALA.
440		420-440	(glauconite) FRANCONIA
	Cambrian		
520			

TABLE 2.—NEW K/AR AGES AND ANALYTICAL DATA

All analyses are on pure biotite fractions. Brackets indicate replicate determinations.

	Age (m.y.)	K (per cent)*	Radiogenic Ar^{40} (ppm)†	Radiogenic Ar^{40} / Total Ar^{40}
Rocky Boy stock, Bearpaw Mtns.	52	7.76	.0292	0.57
			.0290	0.35
Dartmoor granite, Cornwall	290	7.05	.159	0.51
			.157	0.86
Vosges, France:				
Andlau	298	6.29	.146	0.70
		6.41		
		7.44	.180	0.84
Lac Blanc	314	7.45	.180	0.88
		7.34		
Natzviller	324	6.24	.157	0.88
		6.82	.170	0.92
Grosse Pierre	318	6.89	.165	0.85
		6.80		
		6.76		
		7.01	.173	0.78
Chattanooga shale bentonite layer (Late Devonian)	340	5.98	.157	0.73
			.164	0.67
			.155	0.82
Calais, Maine, granites:				
			.158	0.88
			.156	0.51
Red Beach	403	4.83	.153	0.89
			.153	0.79
Keene Lake	407	5.66	.183	0.76
Charlotte	406	5.79	.187	0.89
Meddybemps	404	5.20	.167	0.92
Bessemer, Alabama, bentonite (Middle Ordovician):				
Pyne mine	420	6.94	.232	0.81
			.233	0.90
Wenonah No. 8 mine	419	6.23	.209	0.77
			.207	0.85

* Perkin-Elmer flame photometer analyses by W. W. Brannock and P. L. D. Elmore. $K^{40} = 1.22 \times 10^{-4}$ g/gK

† Isotope-dilution analyses by H. Faul and H. H. Thomas

the Franconia sandstone, are concordant, yet clearly low as compared to the other data in Table 1, and we are faced with the question whether or not glauconite is a suitable mineral for this purpose. Argon and strontium loss must be considered, but the possibility of later overgrowths on glauconite (or some other means of introducing more potassium and rubidium long after the sediment was laid down) seems more likely. Independent evidence for such a process is lacking. Addition of potassium and rubidium in proportion to the amounts already present would produce erroneously low but almost concordant ages.

At the same time we should consider the possibility that the glauconite age is right, and all the "hard-rock" ages are too high because of inherited old radiogenic isotopes (Holmes, private letter, 1958). Inherited daughter products in the minerals from hard rocks would make these minerals seem older than they are, but again it seems improbable that such old radiogenic lead, argon, and strontium could be absorbed in different growing crystals in the proportion that would give concordant ages when now measured. The absence of argon from Paleozoic margarite (Damon and Kulp, 1957) and the concordant Rb/Sr–K/Ar ages measured on very young micas (Jäger and Faul, 1959) would indicate that micas, at any rate, do not usually absorb significant amounts of old radiogenic isotopes when they crystallize.

Construction of a Time Scale

When we now attempt to construct a time scale by reasonable interpolation between these points (Table 1), it becomes obvious that the available data are still too few, too poor, and internally inconsistent. A rough time scale could be constructed on the igneous and metamorphic rocks alone (Mayne *et al.*, 1959; Faul, 1959; Kulp, 1959), and this time scale would be about a third longer than the Holmes-Marble scale (Committee on Measurement of Geologic Time, 1950). Another time scale could be interpolated between the glauconites, and that scale would roughly agree with Holmes-Marble (Hurley, Pinson, *et al.*, 1959; Amirkhanov *et al.*, 1958; Herzog *et al.*, 1958; Lipson, 1956; 1958; Wasserburg *et al.*, 1956). The few data on volcanic tuffs seem to support the hard-rock scale, but the agreement is far from perfect.

Until these difficulties are resolved, any new time scale constructed either on glauconites or on tuffs or on granitic rocks alone is questionable. Extensive research on the subject is in progress. A new time scale will gradually evolve, based on sound geologic ground work, interlaboratory analysis, and broad co-operation among the many interested workers.

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